DENDROCHRONOLOGICAL PERSPECTIVES ON SEASONAL HYDROCLIMATE OF THE COLORADO PLATEAU REGION, USA

by

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As members of the Dissertation Committee, we certify that we have read the dissertation prepared by Rebecca Lynn Brice, titled *Dendrochronological perspectives on drought and hydroclimate extremes of the Colorado Plateau, USA*

and recommend that it be accepted as fulfilling the dissertation requirement for the Degree of Doctor of Philosophy.

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DEDICATION

To women, hold on tight to your dream.

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ABSTRACT

The recent, prolonged drought and increasing aridity with climate change in the Colorado Plateau region has led to dramatic and lasting impacts to water resources and ecosystems, and has led to questions about the viability of native communities in this already arid place. Rising temperatures evaporate moisture from the surface water critical to livestock and municipal supplies, and evaporate water from soil moisture essential for plant growth and water storage. Snow, the dominant source for surface water on the Colorado Plateau, is projected to decline across the western United States with a warmer future. As a consequence of drought, warming temperatures, and changes in winter precipitation, water availability is a critical issue to local communities and to water managers on the Colorado Plateau.

In this dissertation, I consider water resources of the Colorado Plateau in terms of the seasonal climate variables that influence those water resources during the period of instrumental data collection, and over the long-term tree-ring derived reconstructed record. The deficiency of long-term hydroclimatic records in this region precludes deeper analysis of the hydroclimatic relationships affecting water resources on the Colorado Plateau. Limited record length inhibits the ability to evaluate trends and short records fail to capture substantial variation expected in semi-arid regions. A short period of overlap makes it difficult to assess relationships of precipitation, temperature, and streamflow with other variables such as soil moisture and snow water equivalent. Here, I use three aspects of dendrochronology to consider water on the Colorado Plateau, and evaluate the potential to expand dendrochronological science in the region.

The first study investigates the potential for tree rings in the Four Corners to serve as proxies of long-term seasonal climate variability at a regional scale. I evaluated the trees ability to capture monthly precipitation and temperature us-

ing earlywood and latewood ring-widths. I found statistically significant relationships between earlywood- /latewood-widths and season-specific climate variables, and that the earlywood of Douglas-fir and Ponderosa pine exhibit a warm-season temperature signal. The documentation of a temperature proxy embedded in the earlywood of species commonly used in dendrochronological research opens new avenues for climate reconstructions that have immediate implications for climate change adaptation

The second study systematically evaluates the poorly understood relationship between autumn soil moisture and variations in streamflow in the Upper Colorado River Basin. I examined the soil moisture-flow relationship during the instrumental period. Findings suggest that soil moisture has a statistically significant relationship to low streamflow when soil moisture is dry in the previous autumn, and to high streamflow when soil moisture is wet in the previous autumn. Then, I explored the potential for using tree-rings to reconstruct autumn soil moisture in the Upper Colorado River Basin. I found a variable soil moisture-streamflow relationship that could be influenced by prolonged wet or dry conditions, and that is not replicated in the tree-ring derived reconstruction. These results provide new insight into the soil moisture-flow relationship over the instrumental period and may inform future water resources decision-making in the future.

The third and final study was motivated by the Navajo Nation Department of Water Resources water managers questions concerning snowpack and surface waters in the Chuska Mountains. Projected declines in snowpack in the western United States associated with broad-scale anthropogenic warming are tightly linked to surface runoff. This is worrisome for the Navajo Nation who disproportionately rely on snow-fed surface waters of the Chuska Mountains, which provide water for traditional summer grazing and agricultural farms, but that have in recent years begun to go dry. To identify long-term trends and variability in snowpack in the Chuska Mountains, I utilized tree rings to produce two 300-year long snowpack reconstructions for Chuska Mountain snowpack. I compared the two reconstructions in terms of their usefulness for Navajo water managers. In each reconstruction, past snow droughts were evaluated relative to the two most recent severe droughts, the millennium drought (2000 to present), and the 1950s drought. Critically, the 1950s drought, from which Navajo tribal members report experiencing severe impacts to their livelihoods, is rivaled and exceeded by other dry snowpack periods prior to the 20th century. The severity and persistence of past snow droughts, which occurred without anthropogenic forcing, signal potential future declines in Navajo snowpack if warming trends continue.

CHAPTER 1

INTRODUCTION

1. Statement of the Problem

Water is a fundamental component of Earths climate system and is of central importance to human life and activities. Among the most pressing concerns related to future climate change are those attributed to changes in the global water cycle. Streamflow, soil moisture, and groundwater are all fundamental components of the water cycle and are integral components of water resources in semi-arid and arid regions. A changing climate is expected to alter the volume, timing, and recharge of water resources in these drier environments (IPCC, 2014). Even as climate change alters hydroclimate and water resources, ongoing measurements to document these changes are lacking. This is especially true for hydrologic components at finer timescales and that are traditionally more difficult to measure. The result of these compounding effects is substantial uncertainty, now and in the future, about the hydrologic relationships that influence water availability. This kind of uncertainty leads to challenging water management planning in semi-arid environments such as the Colorado Plateau. Water is a scarce resource on the Colorado Plateau, and yet it is critical for ranching, farming, and traditional Native American activities tightly linked to seasonal variations in precipitation. For this reason, communities living on the Colorado Plateau are vulnerable to climate extremes, especially drought, because climate extremes can stress limited water resources in ways that are difficult to anticipate.

Precipitation, temperature, and some streamflow records on the Colorado Plateau establish hydroclimatic variability over the last century. Long-term datasets for these variables allow for the documentation and analysis of basic hydroclimatic processes and their relationships to water resources. However, uneven distribution of climate monitoring stations results in large gaps in hydroclimatic information. Large swaths of the Colorado Plateau also have short records, insufficient to capture the range of extremes expected in topographically and climatologically diverse semi-arid regions such as this. Limited information about a system restricts the ability to assess trends and changes in trends, the ability to fully understand natural variability in a particular place or region, the ability to assess hydrologic relationships, and to do this at various spatial scales which are important for developing effective climate change adaptation strategies. For this reason, the research presented here addresses the problem of limited hydroclimatic information in the context of extreme climatic and topographic variation on the Colorado Plateau with implications for resource management.

2. Background

2.1. The dry climate of the Colorado Plateau

The high Colorado Plateau region is the area roughly centered on the Four Corners where the states of Colorado, Utah, Arizona and New Mexico meet, and which includes portions of the Upper Colorado and Lower Colorado River basins. This physiographic region overlaps with the Colorado River watershed and includes portions of major Colorado River tributaries such as the Gunnison River, the Green River, the San Juan River, and the Little Colorado River. The Colorado Plateau is bounded by the Rocky Mountains to the east and by the Basin and Range topography of the Great Basin to the west. Characteristic topography of the Colorado Plateau includes high plateaus and isolated mountain ranges interspersed with erosional features of varying scales. Topographic complexity is an important characteristic of the Colorado Plateau. In order to capture significant differences in precipitation and the full range of temperatures associated with topography in this region, spatially detailed information is required.

Aridity (when potential evapotranspiration exceeds precipitation in the long-term climatological average) is an important aspect of the Colorado Plateau. The region contains some of the driest locations in the United States (Durrenberger, 1972). Drought (a period of abnormally dry conditions) is a common occurrence in the region. Inadequate precipitation diminishes surface and groundwater recharge, limits discharge from springs, and exacerbates existing constraints on drinking water supplies (Ferguson et al., 2011). During drought, rising temperatures and increased vapor pressure deficits increase evaporation from surface water that is critical to livestock and municipal water supplies (Ferguson et al., 2011; Crimmins et al., 2013). Increased evaporative demand also removes water from the soil essential for plant growth (Adams and Kolb 2005; Williams et al., 2010; Niu et al., 2014; Allen, 2014). Extremes in precipitation and temperature affect drought persistence, timing, recovery rates, and the resulting impacts experienced by local communities (Crimmins et al., 2017).

Drought affects the socioeconomic and environmental stability of communities living on the Colorado Plateau (El-Vilaly et al., 2018) and this is likely to be experienced differently at local scales (Ferguson et al., 2011; Voggesser et al., 2013). Extremely dry conditions lead to dramatic and lasting impacts to water resources, crops, food storage, economic conditions and ecosystem services, and have led to questions about the viability of communities in this already arid place (El-Vilaly et al., 2018; Ferguson et al., 2016). Drought can influence the duration and extent of the dry season, which then influences crop productivity and rangeland forage (Meadow et al., 2013; Ferguson et al., 2011). Intense drought has been affecting the central Colorado Plateau beginning around 1999 (Crimmins et al., 2013; Redsteer et al., 2011) and threatens agricultural systems and ecosystems that are critical to supporting tribal ways of life (Meadow et al., 2013). Concerns about water scarcity have been elevated by the impacts of the prolonged recent drought (Ferguson and Crimmins, 2009), by the increased aridity anticipated due to climate change (Redsteer et al., 2010; Vano et al., 2012; Ficklin et al., 2013; Udall and Overpeck, 2017), as well as the deep cultural connections to a landscape already prone to limited water resources (e.g. Colby et al., 2005; Redsteer, 2011; Voggesser et al., 2013; Chief et al., 2016). Changes to seasonal climate extremes have the potential to decouple close ties between traditional tribal practices and the hydrologic cycle. As a result, there is a need for a more refined and specific understanding of seasonal- to subseasonal-scale hydroclimatic extremes to aid in drought planning and assessment in the region. A deeper understanding of drought and hydrologic response to drought can assist water managers to attenuate negative impacts to future water resources on the Colorado Plateau.

2.2. Hydroclimate on the Colorado Plateau

Hydroclimatology links water resources to the climate system. Precipitation and temperature are the primary climate controls on hydrological variability and therefore drive the quantity and partitioning of water resources. While precipitation amount is the main control of shifts in water balance over space and time, and changes in precipitation have important implications for hydrology and water resources, temperatures also contribute by determining the form of precipitation (rain or snow) and regulating evaporative demand on the land surface. Anomalous precipitation amounts or temperatures can cause water deficits and drought or increased runoff and flooding.

Semi-arid to arid environments experience exceptional variations in precipitation over daily, seasonal, annual and decadal timescales. The semi-arid Colorado Plateau is located at the convergence of two seasonal atmospheric circulations (Schwinning et al., 2008), and as a consequence it receives precipitation split over two seasons, winter/spring and summer (Sheppard et al., 2002). In summer, isolated and unevenly distributed convective storms develop, fed by tropical moisture delivered with the North American Monsoon (NAM) from the Pacific Ocean, the Gulf of California, and the Gulf of Mexico (Adams and Comrie, 1997). Hemispheric-scale circulation modes influential in winter months (December-March) control the delivery of moisture to the region. The winter storm track, guided by the jet stream, drives North Pacific moisture south resulting in widespread cool-season precipitation on the Colorado Plateau (Crimmins et al., 2017; McCabe et al., 1994; Sheppard et al., 2002). The periodic El Nio Southern Oscillation (ENSO) in the equatorial Pacific influences variability in extremes at inter-annual timescales (Cayan et al., 1998), and the Pacific Decadal Oscillation (PDO) is associated with multi-decadal drought cycles (Hereford et al., 2002). Shifts in the Pacific Decadal Oscillation (PDO) and ENSO, and the phasing of these shifts, impacts water resources in the western United States. (Biondi et al., 2001; McCabe and Dettinger, 2002).

Water resources on the Colorado Plateau come predominantly from cool-season precipitation (Sheppard et al., 2002) and are stored in snowpack, springs, groundwater, streams and lakes. Streamflow is the main water supply for the region. The Colorado River dominates this surface hydrology as spring runoff. It originates primarily as snowpack in the high mountains and is directed through the Colorado Plateau while draining approximately 90% of the plateau itself (Patton et al., 1991). Projected declines in snowpack in the western United States are associated with broad-scale warming related to climate change (Knowles et al., 2006; Melillo et al., 2014; Lute et al., 2015; Mankin et al., 2015; Mankin and Diffenbaugh, 2015; Fyfe et al., 2017) and these declines raise concerns about future impacts on water supplies from runoff (Pederson et al., 2011; Redsteer et al., 2013; Garfin et al., 2013; Mote et al., 2018). Snowpack and the water contained in snowpack vary from one year to the next, with runoff linked to these fluctuations (Cayan, 1996; Cayan et al., 1998; Mote et al., 2018).

Although significant long-term trends in annual precipitation in this region are not evident in the historic record, distinct long-term trends in temperature and runoff indicate a continued shift toward streamflow declines with future warming (Nowak et al., 2012; Woodhouse et al., 2016; Xiao et al., 2018). Warm temperatures, changes in winter precipitation from snow to rain, earlier snowmelt, and persistent drought make water availability a critical issue to local communities and to water managers in the region (Garfin et al., 2013).

A range of physical, biological, or human variables may attenuate natural variation in the hydrology of the Colorado Plateau, including soil moisture, snow water equivalent, vegetation cover, and land use patterns. Although runoff and streamflow are the primary focus of water resource concerns on the Colorado Plateau, other components of the hydrologic cycle play a crucial role in water that maintains, replenishes, or moderates Colorado Plateau surface water supplies. Additional components of the hydrologic cycle explored in this dissertation include autumn soil moisture storage and snow water equivalent (SWE).

2.3. Dendrochronological approaches to hydroclimate on the Colorado Plateau

One approach to increase hydroclimatic information used to assess long-term variability and to assist with adaptation to future climatic changes is to employ paleoclimatic data from environmental proxies, such as tree rings. Tree rings can be precisely dated and annually resolved so that each ring represents a calendar year. Dendrochronology, the scientific discipline of using tree rings to study environmental and cultural phenomena, began in the US with observations by the astronomer A. E. Douglass near Flagstaff, Arizona on the southern reaches of the Colorado Plateau (Douglass, 1919). His observations about the changes in annual ring-width patterns, and the consistent presence of these same patterns in cut pine stumps throughout the region, sparked a century of dendrochronological research focused on ring-width variability (Douglass, 1920). Trees record information that can be used to reconstruct climate at annual to seasonal timescales. Because trees can be found in many regions across the globe, they can provide spatially extensive data networks. Tree-ring data from semi-arid regions are calibrated against instrumental records of hydroclimate to develop reconstructions of past climate and to document the occurrence of dry or wet multi-season or multi-year periods, as well as the sequence of seasons and years experiencing moisture extremes (Fritts, 1976; Meko et al., 1995; St. George et al., 2010; Hughes et al., 2011).

Nearly a century of scientific literature documents the hydroclimatic sensitivity of trees on the Colorado Plateau (e.g. Douglass, 1920; Schulman, 1956). Annually-resolved tree rings have been used to reconstruct Colorado Plateau precipitation, temperature, streamflow, and other aspects of hydroclimate (e.g. Woodhouse and Meko 1997; Hirschboeck and Meko, 2005; Salzer and Kipfmueller, 2005; Woodhouse et al., 2006; Meko et al., 2007; Novak, 2007; Anderson et al., 2012; Routson et al., 2011). These paleo records have been applied to analyze the frequency, magnitude, and duration of extreme dry or extreme wet periods, where they have occurred and their spatial extent, and the sequence of seasons and years experiencing these extremes. Tree rings in this region can also be seasonally-resolved using intra-annual tree-ring features, called earlywood and latewood, that reflect cambial development in the early part and late part of the growing season, respectively (Stahle et al., 2009; Griffin et al., 2011; Faulstich et al. 2013). Reconstructions of the seasonal hydroclimatic variables are an important aspect of ongoing research in dendroclimatology (Faulstich et al., 2013; Griffin et al., 2013; Pederson et al., 2011; Woodhouse, 2003; Touchan et al., 2011; Stahle et al., 2009). Seasonal reconstructions hold the potential to provide valuable long-term information about frequency and persistence of water entering, stored in, and leaving the Colorado Plateau.

Dendroclimatological research continues to expand the range of potential information that can be gleaned from tree rings in order to provide information about the past that can inform the present and future. Tree rings can be used to specifically constrain ranges of natural variability more accurately and benchmark high and low extremes. Tree rings also assist in constraining relationships to larger hemispheric-scale ocean-atmosphere dynamics that control hydroclimate. My examination of the Colorado Plateau is inspired by the potential impacts of hydroclimatic extremes, and especially drought, in this region given that impacts of water shortages are already being reported (Redsteer et al., 2011; Ferguson et al., 2016; El-Vilaly et al., 2018; Navajo Nation Water Management Branch, personal communication). My interest is to improve how paleoclimatic science in this region is conducted and to leverage tree rings to do it.

3. Dissertation Outline

In this dissertation, I consider water resources of the Colorado Plateau in terms of the seasonal climate variables that influence those water resources during the period of instrumental data collection, and over the long-term tree-ring derived reconstructed record. The deficiency of long-term hydroclimatic records in this region precludes deeper analysis of the hydroclimatic relationships affecting water resources on the Colorado Plateau. Limited record length inhibits the ability to evaluate trends and short records fail to capture substantial variation expected in semi-arid regions. A short period of overlap makes it difficult to assess relationships of precipitation, temperature, and streamflow with other variables such as soil moisture and snow water equivalent. Here, I use three aspects of dendrochronology to consider water on the Colorado Plateau, and evaluate the potential to expand dendrochronological science in the region.

First, I examine in detail the potential for using more highly refined tree-ring proxies to better capture physiographic complexity found in the Four Corners. A large volume of tree-ring research has been undertaken in the southwestern United States, owing in large part to the history of the science and to the sensitivity of trees to moisture in this region. Opportunities remain, however, to use new and refined aspects of the tree-ring proxy to gain higher specificity in their climate signal. Appendix A documents seasonal precipitation signals contained in sub-annual tree-ring measurements, and reveals temperature signals in the earlywood of commonly used Four Corners tree species. Given future projections of increased temperature and the impacts of elevated temperatures on the Four Corners region, the illumination of a temperature proxy embedded in the earlywood of species commonly used in dendrochronological research opens new avenues for climate reconstructions that have immediate implications for climate change adaptation.

Second, I systematically evaluate autumn soil moisture as a hydrologic component considered to modulate variations in streamflow in the Upper Colorado River Basin (Woodhouse et al., 2016). The Upper Colorado River Basin is another location of extensive tree-ring research, but fine-scale hydrologic relationships between hydrologic components and the Colorado River are largely not well understood. Little is known about the soil moisture and streamflow relationship historically because of the paucity of soil moisture measurements. Appendix B develops a specific understanding of the relationship between soil moisture and Colorado River flow exhibited in the instrumental record, and fits this into the larger hydrologic context. With this understanding, I evaluate the potential for using tree rings to reconstruct autumn soil moisture and I assess if the reconstruction replicates the relationship established in the instrumental record. This study enhances our understanding of hydrologic relationships in the Upper Colorado River Basin, and holds potential to inform future water resources decision-making.

Third, I conduct end user-oriented paleoclimatic research to develop a useful and relevant tree-ring based snowpack reconstruction for the Chuska Mountains, Navajo Nation. Natural climate extremes of the Colorado Plateau, and the impacts of these extremes on the stability of communities living there, underscore the immediate need to engage with local communities and address their local climate questions. The greatest challenge in using tree rings to reconstruct local climate on the Colorado Plateau is short record length. Long-term instrumental records are needed for robust model calibration. A secondary concern is the limited availability of hydroclimatic data at the local scale, especially in regions with sparse spatial coverage of meteorological instrumentation. These temporal and spatial limits further substantiate the need for environmental proxies. Navajo tribal members have reported that snow-fed surface waters in the Chuska Mountains have begun to go dry, and region-wide studies suggest an overall decline in western US snowpack (Fyfe et al., 2017; Mote et al., 2018). But instrumental records of Navajo snow indicate a minor and insignificant decline over the 30 years of the local record (Tsinnajinnie et al., 2018). Given the disconnect between the trend in the limited Chuska snow records and western US snow data, and with the acknowledged importance of surface water to Navajo livelihoods, long-term information about snowpack in the Chuska Mountain is essential to Navajo water managers. I collaborated with the Navajo Water Management Branch in an effort to develop a science product that directly addressed their concerns about Chuska Mountain snowpack variability. Appendix C evaluates this knowledge co-production process and the products of the process, within the context of developing a usable and relevant tree-ring based snowpack reconstruction.

CHAPTER 2

THE PRESENT STUDY

The present study is focused on long-term seasonal information for hydroclimatic elements on the Colorado Plateau. Long-term proxy-based records are necessary to assess regional natural hydroclimatic variability at a variety of time scales. The tree-ring proxy itself can be refined to assist with our understanding of seasonal hydroclimate in complex environments. The methods, results and conclusions of this research are divided into three manuscripts that are included as Appendices to the dissertation. I first investigate the relationship between sub-annual chronologies from moisture-sensitive tree species from the central portion of the Colorado Plateau and seasonal climate variability. I then examine the potential influence of seasonal soil moisture conditions on water-year streamflow in the Upper Colorado River Basin and determine how this relationship differs between the instrumental record and the tree-ring reconstruction. Finally, I leverage tree-growth sensitivity to Colorado Plateau cool-season precipitation in order to produce climate information that is both salient and credible for application by local Navajo water managers.

As allowed by the University of Arizona Graduate College, this dissertation includes three research projects described in three appendices. Each appendix is a co-authored publishable paper and I am the lead and corresponding author for each. The first paper will be submitted to *Tree-Ring Research* (Appendix A), the second paper will be submitted to *Water Resources Research* (Appendix B), and the third paper will be submitted to *Climate Services* (Appendix C). I led the execution of these projects, including data analysis, interpretation of results, and writing, though the final manuscripts are abundantly improved by the support and contribution of co-authors and mentors. Below I summarize the methods and most significant results for each study.

1. Appendix A: Seasonal Climate Signals in Earlywood and Latewood Ring-Width Chronologies of the Four Corners, USA

One goal of this study was to utilize a set of sub-annual tree-ring chronologies for the Four Corners region to determine how these chronologies reflect monthly precipitation and temperature at the seasonal scale. A second goal was to investigate whether sub-annual chronologies can capture nuance in the instrumental record in this physiographically complex region.

This paper asks the following research questions:

- What is the relationship of Four Corners *P. meziesii* and *P. ponderosa* sub-annual tree-ring chronologies to seasonal climate?
- Is there general agreement in the intra-annual climate-growth relationship?
- Do sub-annual tree-ring chronologies capture seasonal temperature?
- What is the potential for seasonal temperature reconstructions?

I used earlywood, latewood, and adjusted latewood ring-width chronologies from moisture-sensitive Ponderosa pine (*Pinus ponderosa*) and Douglas-fir (*Pseudotsuga menziesii*) with monthly PRISM temperature and precipitation to evaluate the climate-growth response of these species in the Four Corners region. Monthly climate signals in the time series of tree-ring indices were identified using the Matlab function seascorr (Meko et al., 2011). The climatic response of the chronologies varied by species and by sub-annual chronology so that seasonal precipitation and temperature signals were distinguishable in earlywood and latewood growth variability. Seasonal precipitation results were consistent with seasonal precipitation reconstructions from sub-annual tree-rings in this region. The earlywood of both species exhibited a warm-season temperature signal that has not before been established using sub-annual ring-widths. This temperature signal expands the potential for seasonal reconstructions of climate in the Four Corners using sub-annual ring-width measurements.

2. Appendix B: Autumn Soil Moisture Reconstruction from Tree Rings in the Upper Colorado River basin, U.S.A

The Upper Colorado River Basin is a highly regulated and engineered watershed. As such, inter-annual streamflow variability is an important aspect of Colorado River water management. Understanding hydrologic components that modulate streamflow variability is an important aspect of this management. Autumn soil moisture is a hydrologic component believed to modulate flow, but is not well understood. Its relationship to streamflow requires systematic evaluation.

This paper asks the following research questions:

- What is the relationship between antecedent fall soil moisture conditions and streamflow in the 100-year-long instrumental record?
- Can antecedent fall soil moisture be skillfully reconstructed utilizing tree rings?
- Does the soil moisture reconstruction replicate the relationships in the instrumental data in the period of overlap?

I used soil moisture output from a hydrologic model (McCabe and Wolock, 2011) and naturalized streamflow for the Colorado River at Lees Ferry to evaluate the soil moisture-streamflow relationship in the instrumental record. Superposed Epoch Analysis (SEA) was used as a statistical compositing method to determine average soil moisture wetness intensity at specified time lags from an anomalous flow year. The frequency of soil moisture conditions in years leading up to an anomalous flow was determined using counts. Initial instrumental analysis using SEA exhibited a strong significant statistical relationship between antecedent soil moisture conditions and anomalous flows in the same water-year and, with lesser significance, one year prior to anomalous flow. Counts of the occurrence of very dry-dry soil moisture conditions in the instrumental record occur most frequently in the same year as an anomalous flow. An important aspect of the instrumental investigation of the soil moisture-flow relationship was the use of linear regression to highlight shared variance between the two records. During antecedent months, soil moisture and water-year streamflow variability have a statistical association with October and November precipitation as well as July and September temperature. These results provide new insight into the soil moisture-flow relationship over the instrumental period and may inform future water resources decision-making in the future.

To develop a November soil moisture reconstruction, I used tree-ring chronologies from the Upper Colorado River Basin as predictors and the soil moisture output from the McCabe and Wolock (2011) hydrologic model as the calibration dataset. SEA, counts, and regression analysis that I used for the instrumental period were replicated for the reconstruction and the results were examined for similarities and differences with the instrumental results. During the period of overlap with the instrumental record (1906-1998), reconstructed soil moisture is not reflecting the same time-dependent relationship. While the instrumental SEA shows the largest average deficit or surplus in soil moisture in the same year as an anomalous flow, the largest average deficit or surplus in reconstructed soil moisture is two years prior to an anomalous flow. Counts of very dry or very wet reconstructed soil moisture occur most commonly in the same year as an anomalous flow during the common period. There is also a higher occurrence of dry soil moisture in the same year as a low flow in the reconstructed period than during the instrumental period. Reconstructed soil moisture does not demonstrate the same capacity to capture moderate conditions relative to an anomalous flow as does the instrumental record. Correlation analysis of reconstructed soil moisture with precipitation and temperature shows that reconstructed soil moisture and reconstructed streamflow share a larger proportion of variance in autumn precipitation and summer temperature than do soil moisture and streamflow in the instrumental record. Inflated shared variance in the reconstruction may have arisen from the use of similar tree-ring chronologies in both models, and which could have artificially inflated relationships between reconstructed soil moisture and reconstructed streamflow. These results demonstrate that the variable soil moisture-streamflow relationship found in the instrumental record may prove difficult to replicated using tree rings as a proxy.

3. Appendix C: Comparing Tree-Ring Based Reconstructions of Snowpack Variability at Different Scales

In order for climate information to be useful and relevant to users, effective co-production of knowledge is required. In the case of Navajo water managers, the focus of this study, there was a need for long-term information about Chuska snowpack, which provides most surface water for the eastern portion of the Navajo Nation and for which there are short instrumental records. Consequently, this study was initiated by water managers with concerns about water resources directly related to Chuska snowpack, the limited information about snowpack amount and variability, and the prospect of ongoing drought and a warming future. This study compares two reconstructions developed through a partnership with Navajo water managers to assess the usefulness and relevance of each reconstruction to the Navajo.

This paper asks the following research questions:

- Can tree-rings be used to reconstruct Navajo snowpack?
- How can end user engagement provide the most useful and relevant climate information for the Navajo Nation?

I used quantitative results from two tree-ring based reconstructions of snowpack to qualitatively assess the ability of these reconstructions to provide useful and relevant climate services to the Navajo Nation Department of Water Resources. The Navajo water managers interest was centered on understanding snowpack variability in their only headwaters in light of a changing climate. Reconstructing snowpack for the Navajo Nation provided unique context, both because of the limited records available to them as well as their considerable vulnerability to hydroclimatic impacts in a warming future. The efforts to develop useful and relevant climate information for the Navajo were tailored to Navajo management practices and targeted to a locally meaningful spatial scale. This research considered how conducting science that is specifically tailored to the needs and goals of the users, can produce useful and relevant information. I experimented with two approaches to answer this question, a reconstruction that was calibrated on limited local snow data and one that was calibrated on a more distal but longer snow record with a strong relationship to the local snow series.

I used tree-ring chronologies collected in northern Arizona on and near the Navajo Nation and two records of snow water equivalent (SWE) reflecting peak snow accumulation to represent local versus regionally representative snowpack. Results show snow droughts to be recurring and persistent in the Chuska Mountains and in the San Francisco peaks west of the Navajo Nation over the last 300 years. Each model skillfully reconstructs SWE, but their comparison reveals differences in the ability of each reconstruction to capture variability present in the instrumental data. Past snow droughts were evaluated relative to the two most recent severe droughts, the millennium drought (about 2000 to present), and the 1950s drought. The local snowpack reconstruction reveals the presence of extremely dry years embedded in longer dry periods, some of which coincide with tangible and documented impacts to Navajo water resources in the 20th and 21st centuries, and do not coincide with other dry periods reflected in the regionally-representative reconstruction. While the reconstruction based upon the representative data was more statistically robust, the more relevant results bore out of the localized reconstruction.

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APPENDIX A

SEASONAL CLIMATE SIGNALS IN EARLYWOOD AND LATEWOOD RING-WIDTH CHRONOLOGIES OF THE FOUR CORNERS, USA

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Abstract

Fourteen earlywood and latewood ring-width chronologies of Douglas-fir (*Pseudotsuga menziesii*) and Ponderosa pine (*Pinus ponderosa*) in the Four Corners region were used to evaluate seasonal climate signals in each of the species-specific chronologies. PRISM data (1895-2018) were compared to each chronology using correlations of monthly precipitation and partial correlations of monthly temperature. Monthly correlations indicate that earlywood of Douglas-fir is significantly correlated with precipitation throughout the cool-season. Differences in monthly precipitation correlations suggest a split climate signal in the earlywood of Ponderosa pine. Seasonal temperature correlations reveal a strong and significant inverse relationship between earlywood of both species in the first months of the warm-season. The climate-growth relationship is more variable in all latewood chronologies. When the influence of earlywood-width on latewood-width is removed, the adjusted latewood seasonal climate signal is dominated by summer precipitation. Seasonal precipitation results are consistent with previous research in this complex study area, but the warm-season temperature signal has not before been demonstrated using sub-annual ring-widths. It may be possible to reconstruct warm season temperature in the Four Corners. With careful site selection and adequate sample depth, earlywood taken from Douglas-fir and Ponderosa pine holds the potential to facilitate warm-season temperature reconstructions in a very diverse study area using a refined measurement of the tree-ring proxy.

Key Words: Douglas-fir, Ponderosa pine, sub-annual, growth response, tree rings.

1. Introduction

Tree species found in the Four Corners region of the southwestern United States such as Ponderosa pine (*Pinus ponderosa*) and Douglas-fir (*Pseudotsuga menziesii*) have been shown for nearly a century to record long-term climate variability in the region (e.g. Douglass, 1919; Fritts, 1976; Dean 1988; D'Arrigo and Jacoby 1991; Salzer and Kipfmueller 2005; Touchan et al. 2011). In semi-arid regions such as the Four Corners, where tree growth is limited by moisture, total-ring width measurements record changes in precipitation that are used to identify and to evaluate years of drought and pluvials on an annual basis over centuries (e.g. Cook et al., 2010; Woodhouse et al., 2010). While annual tree-ring measurements provide long-term records of annual climate variables, tree growth has also been shown to have a relationship with seasonal climate in the southwestern United States, particularly when sub-annual growth increments are considered (e.g. Stahle et al., 2009; Griffin et al., 2011). In the Four Corners, annual totals of precipitation and average temperature obscure seasonal-scale climate variability (Crimmins et al., 2017; House and Hirschboeck, 1993; Sheppard et al., 2002; Fritts, 1976) and sub-annual growth increments serve as promising proxies for examining long-term seasonal climate variability in the region.

The present study builds on previous research in the Four Corners and investigates the potential for sub-annual growth increments in tree rings to serve as proxies of long-term, seasonal climate variability at a regional scale. The sequence of intra-annual growth variation in tree rings is evidence of varying stages of growth as a response to seasonal environmental conditions (Fritts, 2001; Vaganov et al., 2006). These xylogenic variations are species-specific and are exhibited in the cellular elements (e.g. lumen area, cell number, cell wall thickness) in both earlywood and latewood (Ziaco et al., 2014; Fritts, 2001). Light-colored, large and thin-walled cells in conifers indicate earlywood growth and generally reflect environmental conditions leading up to and including the early part of the growing season. Latewood, the darker colored portion of the ring resulting from flattened cells and thick cell-wall development, generally reflects the latter part of the growing season (Hoadley, 1990). Earlywood and latewood ring-widths vary from one year to the next and can be measured separately, resulting in tree-ring chronologies that have the potential to reflect seasonal moisture variability (Cleaveland, 1986; Griffin et al., 2011; Griffin et al., 2013). A network of sub-annual ring-width tree-ring chronologies exists for the Four Corners, including of earlywood (EW), latewood (LW) and adjusted latewood (LWa) indices from two species, *P. ponderosa* and *P. menziesii* (heretofore referred to as PSME and PIPO, respectively). EW measurements from this network effectively capture the winter-moisture signal in the Four Corners region and have been used to reconstruct cool-season precipitation in New Mexico and in the tribal areas of the Four Corners (Stahle et al., 2009; Faulstich et al., 2013). LW growth incorporates the cool-season and warm-season precipitation signals causing statistical inter-correlation between EW-width and LW-width (Meko and Baisan, 2001; Griffin et al., 2011). Meko and Baisan (2001) demonstrated that the summer (July-August) seasonal signal in LW-width measurements can be isolated from EW using linear regression, which removes the statistical dependence of the LW on the EW. In this way, intra-annual tree rings have been used successfully to reconstruct seasonally independent summer precipitation (Stahle et al., 2009; Touchan et al., 2011; Griffin et al., 2013; Faulstich et al., 2013).

The goal of this study was to utilize the existing set of EW and LW chronologies for two species, *P. menziesii* and *P. ponderosa*, to investigate how sub-annual tree-ring widths reflect seasonal climatic information present in the instrumental record of the last 100 years, and to evaluate how refining specific parts of the tree-ring proxy can capture the nuance of more complex environments. Our research questions are: 1) what is the relationship of Four Corners *P. meziesii* and *P. ponderosa* sub-annual tree-ring chronologies to seasonal climate, and 2) is there general agreement among species in the intra-annual climate-growth relationships? We use monthly total precipitation and monthly mean temperature in seasonal correlation analysis with EW and LW chronologies from *P. menziesii* and *P. ponderosa* tree-ring sites. Results of the correlation analysis are then used to evaluate climate-growth relationships within species and between species in the Four Corners region.

2. Study Area

Annual precipitation in the Four Corners region reflects a bi-modal regime (Figure 1d; Sheppard et al., 2002), where precipitation is delivered during two distinct seasons, winter/spring (December-March) and the monsoon (July-September). Seasonal precipitation varies independently so that a wet winter may be followed by a dry monsoon, and vice versa (Faulstich et al, 2013). Precipitation may also vary in concert between seasons so that each season receives above average precipitation or below average precipitation in the same year. These seasonal variations can occur from one year to the next, or for longer periods (Crimmins et al., 2017; Faulstich et al., 2013; Griffin et al., 2013; Touchan et al., 2011; Stahle et al., 2009). The sequence of these seasonal variations has been shown to be critical in defining and understanding impacts of individual droughts on local human systems in the Four Corners (Crimmins et al., 2013; Faulstich et al., 2013; Meadow et al., 2013).

Temperature is a key component in Four Corners climate throughout the year, as a driver of evaporative demand, and is projected to increase in the Four Corners in the future (Crimmins et al., 2017; Williams et al., 2013; Weiss et al., 2009). Aridity in regions such as the southwestern United States is associated with the difference between surface heating and latent heat flux (Sheppard et al., 2002). Warm-season vapor pressure deficit (the difference between the actual- and saturation-vapor pressure largely controlled by temperature) and cool-season precipitation have been shown to be equally important to forest drought-stress in the southwestern United States (Williams et al., 2013). Climate projections of future warming from fifteen

CMIP3 models averaged over the six states of the southwestern United States unanimously show warming in all seasons (Garfin et al., 2013). The ensemble mean-change in air temperature for summer (June-September) over the Four Corners region indicates as much as a 3° C increase in Four Corners summer temperatures over the next 50 years (Garfin et al., 2013). Consensus among climate models for future summer precipitation is considerably weaker than for summer temperatures, though model average precipitation forecasts for spring and summer precipitation in the Four Corners show a decline out to the end of the century (Garfin et al., 2013).

The interval of time between the end of the cool season and preceding the onset of the monsoon in early July is referred to as the pre-monsoon (May-June). The pre-monsoon is the climatological period of lowest precipitation and seasonal warming situated between the cool-season (October-April) and the monsoon (July-August) in the Four Corners (Adams and Comrie, 1997; Grantz et al., 2006). The pre-monsoon is a season when tribal agricultural activities, especially traditional dry land farming, are impacted by variations in the duration and intensity of precipitation and temperature (Navajo Nation Department of Water Resources 2011; Crimmins et al., 2017; Tulley-Cordova et al., 2018).

The Four Corners region is characterized by a complex landscape with high vertical relief (Figure 1a). Low elevations in the area range from about 1,200 m near the entrance of the Grand Canyon at Page, Arizona to nearly 4,000 m at Wheeler Peak, New Mexico. The study area is defined by several mountain ranges: the San Juan Mountains to the north, the Sangre de Cristo and Jemez Mountains to the east, the Zuni and San Mateo Mountains to the south, and the San Francisco Peaks to the west. The central portion of the study area is topographically diverse ranging from deeply incised canyons to high mesas, buttes and mountains. Higher elevations promote forest community development, which correspond, broadly, to the location of mountains and plateaus. Aspect and elevation determine microclimate variations and influence climatic factors most limiting to growth, such as soil moisture and temperature (Burns and Honkala, 1990). Northern aspects and low-grade slopes limit incident radiation, reducing surface heating and soil water loss (Fritts, 1976). Topography, substrate, and orographic site factors can also modify the energy and water balance at a site on seasonal and diurnal time scales (Fritts, 1976). These variations in microsite conditions favor certain conifer species, such as Douglas-fir and Ponderosa pine (Ryker and Losensky, 1983).

Variation in climate generally follows the physiographic elements of the study area in a northeast-southwest gradient. This correspondence is illustrated in four examples: topographic features (Figure 1a), average January temperature (°C; Figure 1b), average July temperature (°C; Figure 1c) and average annual precipitation (cm; Figure 1d). Elevation is an important delineator of temperature variation across the study area. For example, the flat but relatively high (2,300 m) San Luis Valley of south-central Colorado is cool throughout the year, in contrast to the low elevation (1,200 m) western portion of the study area where the Little Colorado River drains into Lake Powell. Average annual precipitation corresponds to elevation, as well. Higher-latitude mountain ranges, such as the San Juan Mountains in Colorado, receive the largest average annual precipitation amounts. This is in contrast with the dry western portion of the study area dominated by lower elevations.

3. Data and Methods

3.1. Climate data

Monthly total precipitation (P) and monthly mean temperature (T) data are from the Parameter-elevation Regressions and Independent Slopes Model (PRISM) gridded dataset (Daly et al., 2008). PRISM data were obtained for the period 1895-2018 from Western Climate Mapping Initiative website (WestMap; data acquired May 2019 from http://www.cefa.dri.edu/Westmap/) within a rectangular area bounded by latitudes 37.25°N and 34.85°N and longitudes -105°W and -

111.75°W (Figure 1). This bounding box encompasses the Four Corners region and includes the tree-ring sampling sites described below. PRISM-derived precipitation and temperature plots illustrate the monthly average range for each climate variable for the Four Corners region. Monthly total precipitation ranges from 1.33 cm in June to 4.12 cm in August. Annual precipitation climatology is dominated by precipitation arriving with the North American Monsoon (NAM) in July and August. Monthly mean temperature averaged across the study area ranges between -1.27°C in January and 22.91°C in July, with summer temperatures at their warmest in June, July and August (1895-2018; PRISM, 2016). As noted above, areas in the Four Corners are relatively wet and others are much drier as a consequence of extreme physiographic variability throughout the region. For example, average annual precipitation in the southwest portion of the study area is about 17 cm and average annual precipitation in the San Juan Mountains is about 136 cm. Temperature similarly varies spatially across the region, between 14°C mean annual temperature in the southwest and 1°C mean annual temperature in the San Juan Mountains.

3.2. Tree-ring data

We examined fourteen tree-ring site chronologies within the Four Corners study region. The tree-ring sites consist of moisture-sensitive tree species, either Ponderosa pine (*Pinus ponderosa*) or Douglas fir (*Pseudotsuga Menziesii*). These data are part of a network of 52 site chronologies collected in the Southwestern United States and Baja, Mexico and were used to generate reconstructions of the North American Monsoon (e.g., Griffin et al., 2013). These data consist of archived chronologies that Griffin et al. (2013) updated in 2009-2011, measured EW and LW ring-widths, and used to develop new chronologies from these measurements. The sub-set of the network used here is comprised of the previously computed residual EW, LW and LWa chronologies selected within a selection-bounding box between 35° to 38° north and 105° to 111.5° west and provided by the original authors (Table 1; Figure

1a). For each site, there are three versions of each chronology, earlywood (EW), latewood (LW) and adjusted latewood (LWa). Adjusted latewood (LWa) is the residuals from linear regression of the measured LW on measured EW in the same year of growth (Griffin et al., 2011). The resultant LWa late summer climate signal is thus independent of the cool season reflected in EW (Meko and Baisan, 2001; Stahle et al., 2009).

3.3. Climatic response analysis

We identified monthly climate signals in the time series of tree-ring indices using the Matlab function of *seascorr* (Meko et al., 2011). One- to twelve- month seasonal correlations were calculated for each climate variable (monthly P and monthly T) for the fourteen months prior to and including the approximate last month of growth (pAugust-September). Simple Pearson correlation, the linear relationship between the climate variable and tree-ring time series, represents the shared variance between climate and chronologies (1895-2008). Partial correlation was used to remove the influence of the first climate variable (P) to determine the relationship between temperature and growth, independent of the relationship with P. *Seascorr* uses Monte Carlo simulation to derive confidence intervals indicating significant correlations and partial correlations following Dietrich and Newsam (1997). Temporal stability was assessed using a difference of correlations test (Snedecor and Cochran, 1989). This test utilizes Fishers Z transformation of correlations for significance and is applied within *seascorr* to the highest-correlated seasons in two sub-periods (early N = 56; late N = 56) and is adjusted for effective sample size (Meko et al., 2011). The difference of correlations test assumes that the time series being correlated are normally distributed. Months of significant correlations were plotted by species chronology, and ordered by elevation.

4. Results

4.1. Monthly climate correlations with Four Corners tree rings: PSME

Each PSME EW chronology is significantly ($p < 0.05$) correlated with precipitation in the months of pOctober through May (Figure 2a), with the exception of the lowest-elevation PSME EW site (WCM), which is significantly correlated through June. Mid-elevation sites (PCM, EAM, DCM) are significantly correlated with pSeptember. High elevations sites are significantly correlated with pAugust (TCM, SPM) and pSeptember (SPM). Table 2 shows the highest monthly correlation coefficients with climate variables for all PSME EW and LWa chronologies. The highest correlation among all PSME EW chronologies is with pOctober and pDecember precipitation ($r = 0.45$). Monthly PSME EW precipitation correlation coefficients range between $r = 0.28$ in March (DCM) to $r = 0.45$ in pOctober (WCM) and pDecember (MVM).

All PSME EW chronologies are significantly ($p < 0.05$) partially correlated with temperature in the months of June and July. PSME EW is also significantly partially correlated with April temperature, with the exception of two mid-elevations sites (PCM and DCM). Two PSME EW sites (WMF and DCM) show a strong temperature correlation with the month of May. Monthly PSME EW temperature correlation coefficients range between $r = -0.33$ in June (TCM) to $r = -0.41$ in May (DCM).

All PSME LW chronologies, with the exception of DCM, show a strong and significant ($p < 0.05$) relationship with precipitation in pDecember and June (Figure 2b), but the cool-season signal present in the EW is also retained in many of the PSME LW chronologies. PSME LW correlations with precipitation in pDecember strengthen over PSME EW correlation in some chronologies (e.g. WCM, PCM, EAM, MVM, TCM, SPM). Sites DCM and WMF LW show a strong correlation with precipitation in July and correlations of these chronologies with months preceding

July are considerably weaker than that exhibited in PSME EW. Previous autumn through early summer precipitation (pOctober to June) is significantly ($p < 0.05$) correlated with the three highest elevation sites, and two mid-elevation sites (PCM and EAM).

Partial correlations with temperature and PSME LW occur mostly in June and July, but these correlations weaken from EW correlations with some chronologies (DCM and WMF) while others strengthen in June and July (e.g. WCM, MVM). There are partial correlations in May with mid- to high- elevation PSME LW chronologies. Partial correlations of temperature with PSME LW diminish in pAugust and pNovember from PSME EW partial correlations with temperature.

When the latewood is statistically isolated from the PSME EW (PSME LWa), the dominant signal is summer precipitation (June-July) (Figure 2c). PSME LWa correlates significantly ($p < 0.05$) at all sites in July, with strongest correlations at two mid- to low-elevation sites (DCM and WMF). Low-elevation sites (WCM, WMF, PCM) and one higher elevations site (TCM) also show significant correlation in pDecember. One-month PSME LWa-precipitation correlation coefficients range between $r = 0.24$ in June (EAM) to $r = 0.48$ in July (DCM) (Table 2). Inverse correlations with pAugust precipitation are also present in PSME LWa. Monthly PSME LWa temperature partial correlation coefficients range between $r = -0.13$ in June (EAM) to $r = -0.30$ in May (WMF). Partial correlations with PSME LWa and temperature are varied but they do demonstrate agreement among the chronologies in the months of May and June.

4.2. Monthly climate correlations with Four Corners tree rings: PIPO

PIPO EW is significantly ($p < 0.05$) correlated with precipitation in the months of pOctober through July (Figure 2d). The highest monthly correlation coefficients with climate variables are shown in Table 3 for all PIPO EW and LWa chronologies. Most of the significant monthly correlations occur in spring through early summer

(March-July). The strongest site correlations with precipitation are low-elevation sites (WMP and TSM) in pDecember and the high-elevation site (VPU) in May. Generally early winter pNovember-January precipitation correlations are stronger at low-elevation sites. Generally spring precipitation correlations (May-June) are stronger at mid- to high-elevation sites. Monthly PIPO EW precipitation correlation coefficients range between r = 0.29 in pNovember (GPU) to r = 0.49 in pDecember (TSM).

Each PIPO EW chronology is significantly ($p < 0.05$) partially correlated with temperature in June and July. Most chronologies also show correlation in pNovember (VPU, SFK, RPU, TSM). Partial correlations with temperature are present in May at low-elevation sites (e.g. WMP, TSM, RPU). The WMP EW chronology is significantly correlated with spring and summer months (April-July). Monthly PIPO EW temperature correlation coefficients range between r = -0.22 in July (GPU) to r = -0.45 in June (WMP).

PIPO LW shows a distinct separation between seasonal precipitation signals. Significant correlations with precipitation are strongest across most of the chronologies in July. TSM is not as strong as the others, though it is still significant, and GPU is not significantly correlated with precipitation in this month. Low- to mid-elevation PIPO LW chronologies correlate with precipitation in winter months as well (pNovember-January) (Figure 2e). TSM shows the strongest significant correlation in these months. Mid-elevation chronologies correlate with April-June precipitation in PIPO LW, and July-August demonstrate the most agreement in precipitation correlations.

Temperature partial correlations with PIPO LW occur mostly in June, and these correlations weaken from EW correlations among all chronologies. Mid-to low-elevation sites (SFK, RPU, TSM, WMP) show late spring partial correlations with temperature. Positive partial correlations between PIPO LW and March temperature are present at high-elevation chronologies (GPU and VPU). The pNovember temperature signal of PIPO EW is not present in PIPO LW.

When the latewood is statistically isolated from the PIPO EW (PSME LWa), the dominant signal is summer precipitation (July-August) (Figure 2f). PIPO LWa correlates strongly and significantly ($p < 0.05$) at all sites in July, with the exception of GPU. High-elevation sites (GPU and VPU) exhibit a negative significant correlation with precipitation in February. Monthly PIPO LWa-precipitation correlation coefficients range between $r = -0.21$ in February (GPU) to $r = 0.5$ in July (WMP). Partial correlations with PIPO LWa and temperature are vastly diminished and demonstrate no substantive agreement among the chronologies. One-month PIPO LWa precipitation correlation coefficients range between $r = -0.13$ in April (WMP) to $r = -0.25$ in September (VPU). It is notable that the high-elevation sites do agree with positive and significant partial correlation with temperature in March

5. Discussion and Conclusions

5.1. Climate signals in the Four Corners chronologies

Significant monthly correlations vary by species and by chronology in the Four Corners. Results from this study demonstrate that seasonal precipitation and temperature signals can be distinguished in EW/LW growth variability. The significant cool-season precipitation signal in the earlywood of *P. menziesii* shown here is consistent with previous studies from the region (e.g. Faulstich et al., 2013; Cleaveland et al., 2003; Meko et al., 2013; Stahle et al., 2009; Villanueva-Diaz et al., 2007). Predictors for the tree-ring derived reconstruction of cool-season precipitation (Oct-Apr) for the Four Corners from Faulstich et al. (2013) are among the *P. menziesii* chronologies in this study with significant correlations to the cool season (SPM and WCM). Our correlation results indicate that the early winter precipitation signal is strongest in *P. menziesii* EW. The *P. menziesii* latewood growth association with antecedent EW accentuates this mix of cool-season and warm-season climate signals (Griffin et al., 2009; Stahle et al., 2009). The cool-season precipitation signal is less consistent in the individual *P. menziesii* latewood chronologies; generally correlations with precipitation in January-March weaken, though the split-season association is stronger in some chronologies. The same is true for spring and early summer, suggesting that LW is capturing the mixed precipitation signal. The *P. menziesii* adjusted latewood results in this study are consistent with previous reconstructions of North American Monsoon precipitation in the southwestern US (Griffin et al. 2013; Faulstich et al., 2013) and warm-season precipitation in the broader NAM monsoon region (Meko and Baisan, 2001; Stahle et al., 2009; Therrell at al., 2002).

When compared to *P. menziesii*, the EW of *P. ponderosa* demonstrates a more pronounced separation between cool-season and warm-season precipitation signals in Figure 2, although the highest monthly correlations shown in Table 3 do not highlight this EW seasonal difference in the same way. The precipitation signal in the seasonally-mixed LW of *P. ponderosa* separates the cool-season influence on LW growth from the warm-season with a distinct break in significant precipitation correlations in February through March. These results also demonstrate a strong and significant relationship between the adjusted latewood of *P. ponderosa* and late summer precipitation (July-August).

The warm-season temperature signal based on monthly correlations documented in this study is notable, and a new contribution to understanding the growth response in EW and LW to seasonal temperature for these two Four Corners species. Warm-season (April-September) temperature has been previously reconstructed in the western US using densitometric tree-ring data from 70 sites and several different species (Briffa et al., 1992). The Southwest Deserts region of the study discussed by Sheppard et al. (2002) shows that tree rings can be used to reconstruct periods of warm-season variation in the southwestern U.S. Here the warm season temperature signal is strongest and most coherent in the earlywood and latewood for *P. menziesii* in April through July, and the earlywood of *P. ponderosa* during May through June.

5.2. Concluding remarks

This study finds that sub-annual earlywood and latewood chronologies from two species in the Four Corners contain complex seasonal climate signals. Distinct differences are present between species, between same-species earlywood and latewood, and between same-species tree-ring sites. Despite the variation in the strength of the monthly climate signal between chronologies, seasonal precipitation was distinguishable in the earlywood and latewood of *P. menziesii* and *P. ponderosa*. The results presented in this study support previous research using seasonal climate signals and sub-annual ring-widths to reconstruct cool-season and warm-season precipitation. The presence of a cool-season precipitation signal in both earlywood and latewood was identified in nascent dendrochronological research from the American southwest (e.g. Fritts, 1976) and in more recent hemispheric-scale work (St. George and Ault, 2014).

Precipitation is variable among chronologies, but warm-season temperature (June-July) in earlywood is significant despite weak correlation coefficients and is consistent across all sites and across the two species (Figure 3). This warm-season signal has not before been demonstrated using sub-annual ring-widths in this region. This result supports the possibility of developing warm-season temperature reconstructions in a very diverse study area using a refined measurement of the tree-ring proxy. The inter-correlation of precipitation with temperature in southwestern U.S. tree physiology challenges the prospect of using moisture-limited trees to reconstruct warm-season temperature, however (Fritts, 1976). Partial correlation analysis facilitates the examination of an independent temperature variable, but the climate-tree growth relationship examined here is likely related to drought stress (a combination of limited precipitation and increased vapor pressure deficit related to higher temperatures) rather than just warm-season temperature as the limiting factor (Williams et al., 2013). To reconstruct warm-season temperature independent of precipitation in this region will require further study.

The consistency of the warm-season signal also simplifies study design and field collection in a complex environment. Earlywood measurements can be taken concurrently with total-ring measurements, thereby limiting expense associated with other intensive methods used to reconstruct temperature. Given the importance of temperature in the Four Corners, and the projected increase in regional temperatures in the future, a consistent and reliable proxy for warm-season temperature reconstructions is vital.

It has been shown in other research that physiographic heterogeneity in the northern Rocky Mountains influences the *P. menziesii* climate signal at the site level (Crawford et al., 2015). Great Basin research on climate-growth relationship at the cellular level shows that the effects of reduced water availability seems more relevant to some species at lower elevations than temperature (Ziaco et al., 2014). The interplay of various factors in the spatially complex Four Corners may be contributing the seasonal precipitation signal variation observed in this study. Due to the physiographic complexity in the Four Corners, factors such as elevation, aspect and exposure to surface heating associated with each chronology are most likely interacting to influence climate-growth relationships.

The differential responses of each species to their environment provide impetus for further research to examine all potential factors including many that were not within the scope of this study (e.g. aspect, site-level microclimate, within-stand competition, age stratification) in order to discern external influences on earlywood and latewood climate-growth relationships in the Four Corners. Such endeavors will require intensive metadata collection. Future analysis should also include evaluation of multiple months (as seasons) to further refine the sub-annual ring-width proxy. Monthly correlations are useful in initial analysis, but the highest monthly correlation values for all chronologies may be similar from one month to the next within a season. If monthly correlations are evaluated independently the results may be misleading, and may contribute to difficulty identifying clear patterns between climate and other environmental influences on growth. Results may further be refined with additional evenly dispersed site collections across the study area. This study demonstrates the potential to use new approaches to tree-ring research that will help gain higher specificity and answer deeper questions at the regional scale.

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Tables

Table 1 Tree-ring site chronologies in the Four Corners region. Each site is comprised of three index chronologies: earlywood (EW), latewood (LW) and adjusted latewood (LWA).

Site Code	Site Name	Species	Inner Year	Outer Year	Latitude (°N)	Longitude (°W)	Elevation (m)
VPU	Vera Platt Bradbury	PIPO	1661	2008	37.470	-106.300	2835
GPU	Garcia Park	PIPO	1879	2008	36.361	-105.399	2718
SFK	South Fork	PIPO	1566	2008	37.660	-106.660	2591
RPU	Rio Pueblo	PIPO	1541	2008	36.159	-105.594	2455
TSM	Turkey Springs	PIPO	1595	2008	35.399	-108.527	2374
WMP	Walnut canyon Pine	PIPO	1528	2010	35.174	-111.515	1995
SPM	Satan Pass Fir	PSME	1313	2008	35.606	-108.131	2306
TCM	Tsegi Canyon	PSME	863	2008	36.681	-110.535	2068
MVM	Mesa Verde Fir	PSME	480	2008	37.170	-108.520	2042
DCU	Ditch Canyon Fir	PSME	1610	2008	36.996	-107.801	2036
EAM	Echo Amphitheater	PSME	1295	2008	36.357	-106.526	2029
PCM	Pueblita Canyon	PSME	1643	2009	36.701	-107.320	2005
WMF	Walnut Canyon Fir	PSME	1645	2010	35.174	-111.515	1995
WCM	White Canyon	PSME	1347	2009	37.620	-109.997	1895

Table 2 One-month seasonal correlation and partial correlation coefficients of PSME EW and LWa with precipitation and temperature. The months listed are those with the highest correlation to precipitation (left month column) and to temperature (right month column). Top panel is EW site chronologies. Bottom panel is LWa site chronologies. Chronologies are ordered top to bottom from low elevation to high elevation.

PSME EW Chronology	Month	Correlation Coeff. with P	Month	Partial Correlation Coeff. with T
WCM	pOct	0.45	Jun	-0.35
WMF	pDec	0.43	Jun	-0.40
PCM	Apr	0.40	Jul	-0.35
EAM	May	0.39	Jul	-0.39
DCM	Mar	0.28	May	-0.41
MVM	pDec	0.45	Jul	-0.38
TCM	pDec	0.40	Jun	-0.33
SPM	pNov	0.42	Jul	-0.38

PSME LWa Chronology	Month	Correlation Coeff. with P	Month	Partial Correlation Coeff. with T
WCM	May	0.40	Jun	-0.26
WMF	Jul	0.48	May	-0.30
PCM	Jun	0.39	Jun	-0.19
EAM	Jun	0.24	Jun	-0.13
DCM	Jul	0.46	May	-0.22
MVM	Jun	0.32	Jun	-0.17
TCM	Jun	0.27	May	-0.27
SPM	Jun	0.38	May	-0.17

Table 3 One-month seasonal correlation and partial correlation coefficients of PIPO EW and LWa with precipitation and temperature. The months listed are those with the highest correlation to precipitation (left month column) and to temperature (right month column). Top panel is EW site chronologies. Bottom panel is LWa site chronologies. Chronologies are ordered top to bottom from low elevation to high elevation.

PIPO EW Chronology	Month	Correlation Coeff. with P	Month	Partial Correlation Coeff. with T
WMP	pDec	0.42	Jun	-0.45
TSM	pDec	0.49	Jun	-0.37
RPU	pNov	0.34	Jul	-0.30
SFK	May	0.33	Jun	-0.35
GPU	pNov	0.29	Jul	-0.22
VPU	May	0.40	Jun	-0.30

PIPO LWa Chronology	Month	Correlation Coeff. with P	Month	Partial Correlation Coeff. with T
WMP	Jul	0.50	Apr	-0.13
TSM	Aug	0.44	Jul	-0.13
RPU	Jul	0.45	Sep	-0.20
SFK	Jul	0.46	May	-0.15
GPU	Feb	-0.21	Sep	-0.14
VPU	Jul	0.29	Sep	-0.25

Figures

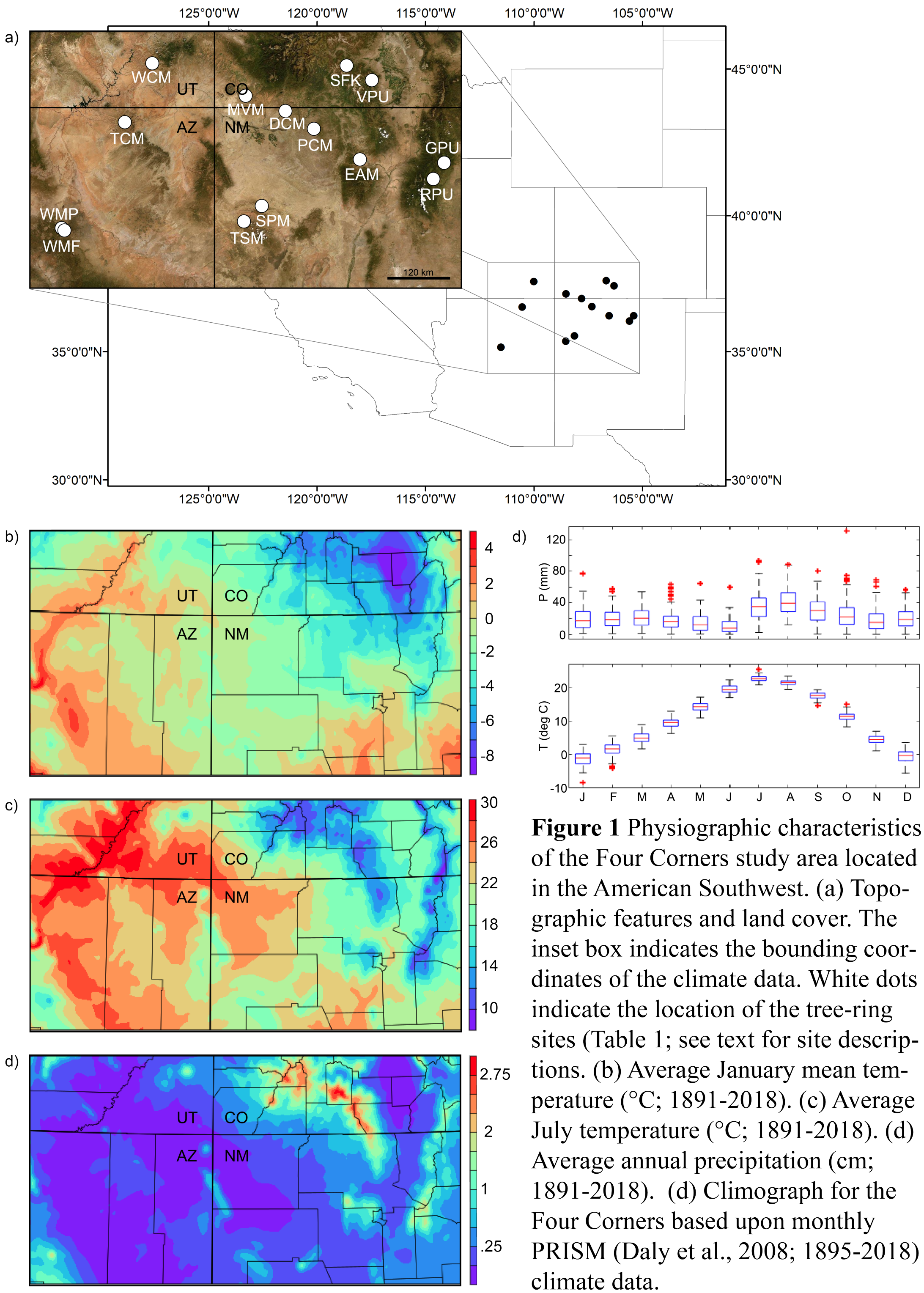

Figure 1 Physiographic characteristics of the Four Corners study area located in the American Southwest. (a) Topographic features and land cover. The inset box indicates the bounding coordinates of the climate data. White dots indicate the location of the tree-ring sites (Table 1; see text for site descriptions. (b) Average January mean temperature (°C; 1891-2018). (c) Average July temperature (°C; 1891-2018). (d) Average annual precipitation (cm; 1891-2018). (d) Climograph for the Four Corners based upon monthly PRISM (Daly et al., 2008; 1895-2018) climate data.

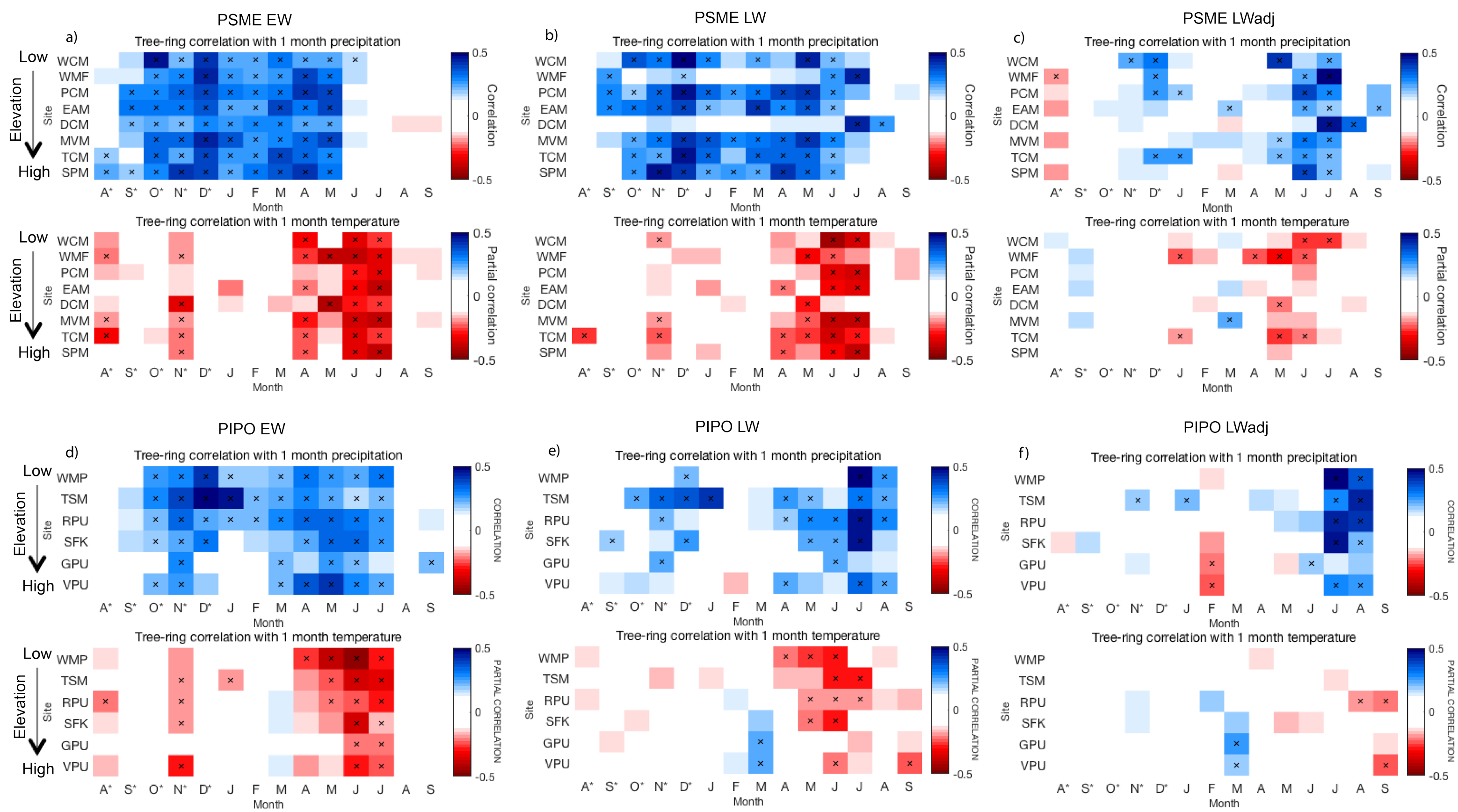

Figure 2 One-month correlations (P) and partial correlations (T) of monthly climate variables (P/correlations top plot, T/partial correlations bottom plot in each panel) and tree-ring site chronologies shown at left of each plot. a) Douglas-fir earlywood, b) Douglas-fir latewood, c) Douglas-fir adjusted latewood, d) Ponderosa pine earlywood, e) Ponderosa pine latewood, and f) Ponderosa pine adjusted latewood. Tree-ring sample sites are ordered from low (top) to high (bottom) elevation. Correlation (P) and partial correlation (T) coefficients are indicated with the color bar at right of each plot. Months are shown as letters along the bottom of each plot, the asterisk indicates the year before. Significant correlations ($p < 0.05$) denoted with X.

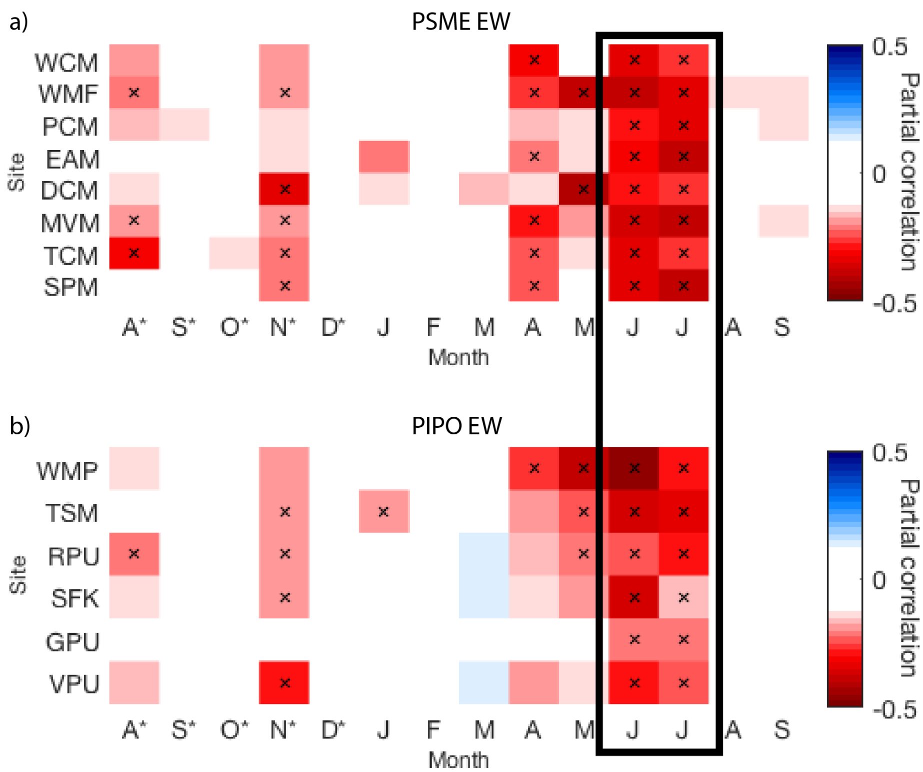

Figure 3 Earlywood partial correlations with temperature at all sites and for both species, a) Douglas-fir, and b) Ponderosa pine. The black box highlights the consistent significant ($p < 0.05$) relationship with warm-season temperature.

APPENDIX B

AUTUMN SOIL MOISTURE RECONSTRUCTION FROM TREE RINGS IN THE UPPER COLORADO RIVER BASIN, U.S.A

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Abstract

Seasonal contributions of precipitation, temperature, soil moisture, and base flow influence streamflow variability in the Colorado River basin, where streamflow variability is an important aspect of Colorado River water management. Autumn soil moisture has the potential to moderate snowmelt runoff in the spring, influencing already low or high flows in the Colorado River. Therefore, it is important to understand the association between autumn soil moisture and subsequent streamflow. Here, we use soil moisture output from the McCabe and Wolock (2011a) hydrologic model and Colorado River naturalized streamflow data to evaluate the relationship between autumn soil moisture and anomalous flow in the Colorado River over the 100-years of instrumental record. We then use soil moisture with moisture-sensitive tree-ring chronologies to reconstruct November soil moisture over the last 513 years in the Upper Colorado River basin. The reconstruction is used to investigate the ability of tree rings to replicate the relationship of November soil moisture to low or high water-year flows evident in the instrumental record. We find that significantly dry/wet November soil moisture conditions in the instrumental record occur in the same year as, and one year prior to, an anomalous low/high flow, respectively. In the reconstructed common period (1906-1998), the largest average soil moisture deficit/surplus is two years prior to a low/high flow. Counts analysis of the reconstructed common period reveals very dry or very wet soil moisture occurs most commonly in the same water year as a low or high flow, respectively. We found higher occurrence of dry soil moisture in the same year as a low flow in the Colorado River over the reconstructed common period than during the instrumental analysis, and in comparison to the wet soil moisture-high streamflow relationship. In most cases, soil moisture deficit/surplus is consistent in intervening years when a lagged relationship is present. Our results suggest a variable soil moisture-streamflow relationship in the instrumental record that could be influenced by prolonged wet or

dry conditions and that may prove difficult to replicated using tree rings as a proxy. Refinement of the autumn soil moisture-streamflow relationship may enhance our understanding of hydroclimatic variability in the Colorado River basin and inform future water resources decision-making.

Key Words: Colorado River Basin, soil moisture, water year streamflow, tree rings

1. Introduction

Fluctuations in Colorado River streamflow reflect changes in runoff driven primarily by precipitation, and to a lesser extent temperature (Nowak et al., 2012; Vano et al., 2012; Woodhouse and Pederson, 2018). While the highly regulated Colorado River has been engineered to buffer the impacts of Upper Colorado River Basin (UCRB) drought through a large amount of storage, interannual streamflow variability is still an important aspect of Colorado River water management (Jerla et al., 2012). Recent concerns about the reliability of Colorado River supply to meet current and future water demands have prompted research on the relationship between climate and streamflow (e.g. Cayan et al., 2010; Vano et al., 2012; Vano et al., 2014; Castle et al., 2014; Woodhouse et al., 2016; McAfee et al., 2017; Udall and Overpeck, 2017; Xiao et al., 2018; Woodhouse and Pederson, 2018).

Streamflow variability in the UCRB is influenced by seasonal hydroclimate including winter snowpack, spring temperatures, and autumn soil moisture (Woodhouse et al., 2016). Woodhouse et al., (2016) found that cool season precipitation (October-April) is the most important variable for streamflow in the UCRB, accounting for 66% of the variance in water-year flow. Warm temperatures have also been shown to reduce runoff efficiency and play an important role in moderating streamflow in certain conditions (Harding et al., 2012; Woodhouse et al., 2016; McCabe et al., 2017; Udall and Overpeck, 2017; Xaio et al., 2018).

Some studies suggest that soil moisture (the amount of water stored in the soil) can influence runoff and water-year streamflow as well, moderating or exacerbating the impacts of high or low flows (Hamlet et al., 2007; Hawkins and Ellis, 2010; Anderson et al, 2012; Mahanama et al., 2012; Rosenberg et al., 2013; Woodhouse et al., 2016). Soil moisture influences the partitioning of precipitation between infiltration (the entry of liquid water into soil) and surface runoff. After infiltration, precipitation is lost through evapotranspiration as the water in the soil is transferred from the soil and vegetation to the atmosphere. These processes integrate the precipitation inputs and losses in soil moisture (Hamlet et al., 2007; Seneviratne et al., 2010; Zhu et al., 2014). Soil moisture has been shown to be an especially important control on infiltration and runoff in snow-dominated mountainous regions in the western US and elsewhere (Hamlet et al., 2007; Penna et al., 2010; Huntington and Niswonger, 2012). Soil moisture conditions in autumn preceding snowpack accumulation (antecedent soil moisture) especially at higher elevations, are believed to influence runoff efficiency, the ratio of annual runoff volume to total precipitation volume in a basin (Burnash et al., 1973). Because soil moisture has been found to influence runoff efficiency, the Colorado Basin River Forecast Center uses soil moisture as part of water supply forecasts and water allocation estimates for the Colorado River (CBRFC; Burnash et al., 1973; see also https://www.cbrfc.noaa.gov/wsup/sac sm/sacsm.php).

Autumn is a critical period for conditioning soil moisture for spring snowmelt in the UCRB (Miller et al., 2014) and may be used to anticipate soil moisture influence on streamflow in the following year (Hawkins and Ellis, 2010). Here we illustrate the infiltration process, starting in the fall, using a conceptual model (Figure 1). For most of the UCRB, as temperatures decline in late fall, water contained in the soil matrix will freeze for winter. At the time of freeze, soil conditions and snow cover are important for determining if water will preferentially infiltrate during snowmelt or, if infiltration is restricted by ice, snow melt will run off (Iwata et al., 2010). If the soil is saturated at the time of freeze, ice forms in the large pores of the soil matrix and slows or stops water movement through the profile. If the soil is unsaturated at the time of freeze, the largest pores are air-filled and infiltration potential is increased (Mohammed et al., 2018). As snowmelt begins in spring, the state of soil conditions in the fall will be critical to infiltration-runoff partitioning of the snowmelt. Although the soil is still frozen as the snowpack warms and reaches wetting and coarsening within the snow grains in preparation for melt, if the soil matrix has greater porosity because it is not congested with frozen water there are pathways for melt water at the snowpack-soil interface and below (Figure 1; Viessmann and Lewis, 2003). In years when the soil moisture conditions in the previous late autumn (November), and at the time of soil freeze, are dry there will be fewer frozen pores in the unsaturated zone (Figure 1a, b). For this reason, as snow melts the water will preferentially infiltrate the unsaturated reservoir to fill open pores first, reduce runoff efficiency, and will be evident in lower-than-expected stream flows given the water contained in the snowpack (Figure 1c, vertical arrows indicate infiltration amounts relative to wet, average and dry snowpack years and angled arrows represent relative runoff). In years when the soil moisture conditions in the previous late autumn and at the time of soil freeze are wet (Figure 1d, e) less snowmelt will enter the soil. Runoff will be higher in these years, resulting in higher-than-expected streamflow given the water contained in the snowpack (Figure 1f).

Because of the role that antecedent soil moisture plays in infiltration of snowmelt, quantifying variability in autumn soil moisture is critical for understanding the relationship between soil moisture and spring runoff. However, actual records of measured soil moisture are few. There are 13 soil moisture measurement SNOTEL stations in the UCRB. The longest of these at the time of this study were less than 10 years, too short to make an assessment of relationships between streamflow and soil moisture. In an attempt to circumvent the paucity of in-situ soil moisture measurements, long-term soil moisture is often approximated using an index such as the Palmer Drought Index (e.g. Palmer 1965; Perry and Niemann, 2007; Snchez et al., 2016; Halwatura et al., 2016) or hydrologic models based on historical measurements of individual hydrologic components (e.g. Fan and van den Dool, 2004; McCabe and Wolock, 2011a). These models estimate hydrologic conditions including evapotranspiration, snow melt, and water contained in the soil using precipitation, temperature and other variable inputs (Figure 2; McCabe and Wolock, 2011a). Output from such models can be used to better understand the influence of autumn soil moisture conditions on water-year streamflow in the instrumental record.

Given the scarcity of soil moisture data, and the potential influence of autumn soil conditions on water year flow, we evaluate the soil-flow relationship in the instrumental data. There is also a need to develop a better understanding of the full natural variability intrinsic to soil conditions and their relationship to runoff. Short soil moisture records limit the range of variability captured by the data and preclude the use of more than one instrumental record to evaluate past hydrologic relationships. Environmental proxies can extend the soil moisture record beyond the instrumental period, enabling a deeper understanding of natural variability in soil moisture patterns and facilitating long-term comparisons to other reconstructions of hydroclimatic variables. Tree-rings have the potential to provide useful proxy information because of the well-understood tree-growth response to moisture, especially in arid and semi-arid environments such as the UCRB (Meko et al., 1995). Trees that grow in these environments are often limited by the amount of water available in the soil and the evaporative demand from the atmosphere, which influences ring-width variation from one year to the next.

A large body of work has shown that tree rings can be used to reconstruct a diverse set of hydroclimatic variables in the UCRB. The pioneering work of Stockton and Jacoby (1976) used tree rings to demonstrate that the Colorado River streamflow record could be extended back in time, allowing the droughts and pluvial periods of 20th century to be evaluated in a long-term context. Subsequent reconstructions of Colorado River streamflow developed using different statistical approaches and a large network of tree ring data underscored tree-rings as a skillful proxy, explaining as much as 81% of the variance in observed streamflow records (Hidalgo et al., 2000; Woodhouse et al., 2006; Meko et al., 2007). Tree rings have been used to produce highly skillful reconstructions of other hydroclimatic variables such as snowpack (Pederson et al., 2001; Woodhouse et al., 2003; Timilsena and Piechota, 2008), the Palmer Drought Severity Index (PDSI; Cook et al., 2004) and runoff efficiency (Woodhouse and Pederson, 2018). Summer soil moisture storage and potential evapotranspiration are shown to be possible as well (Gangapadhyay et al., 2015).

Tree growth has been shown to be closely associated with soil moisture suggesting that, with a sufficient period of overlap, tree-ring growth may be employed to estimate soil water availability (e.g. Bassett, 1964; Cook et al., 2007; St George et al., 2010; Anderson et al., 2012; Griffin and Anchukaitis, 2014). Anderson et al. (2012) demonstrated the potential for using tree rings to reconstruct annual soil moisture in the UCRB using modeled soil moisture rather than a soil moisture index, such as the Palmer Drought Severity Index (PDSI), which mirrors but does not precisely quantify soil moisture. Anderson et al. (2012) also demonstrated that soil moisture and other hydrologic variables are highly correlated, suggesting that the strong connection between tree growth and hydrologic variables can be leveraged with soil moisture as well. Gangapadhyay et al. (2015) demonstrated that tree-ring based reconstructions (1404-1905) of soil moisture storage and other hydroclimatic variables can be generated for the UCRB using non-parametric reconstruction techniques and a water balance model. Building on the work of Anderson et al. (2012) and Gangapadhyay et al. (2015), we assess the potential for utilizing tree rings to gain a long-term perspective on autumn soil moisture conditions rather than annual or summer (Anderson et al., 2012; Gangapadhyay et al., 2015).

The integrating characteristic of soil moisture and its relationship to hydrology in the UCRB means that tree-ring proxies could provide long-term records that illuminate hydroclimatic relationships not apparent during the shorter period of instrumentation. For instance, previous research indicated that the 20th century has been relatively wet in the UCRB (e.g. Stockton and Jacoby, 1976; Woodhouse et al., 2006; Pederson et al., 2011) and that warmer and drier periods have occurred in the basin in earlier centuries, contributing to our understanding of the range of natural hydrologic variability possible and with implications for water resource management (e.g. Stockton and Jacoby, 1976). Increased information about the contribution of antecedent soil moisture to anomalous Colorado River flows may guide water mangers and users future expectations about water resources. For this reason, it is important to gain a longer-term understanding of autumn soil moisture conditions and their relationship to annual water-year Colorado River streamflow during the instrumental period and in the paleo record.

The relationship between autumn soil moisture and UCRB water-year flow has not yet been evaluated, and the ability of tree rings to replicate this relationship has not yet been demonstrated. Therefore, the relationship in the instrumental record must first be established, and then the reconstructed relationship may be compared with that in the instrumental record. This comparison is a critical step for two reasons. First, since 20th century hydroclimate is not an exact analog for previous centuries, the instrumental record is expected to reflect hydroclimatic relationships unique to the 20th century. Second, if the trees cannot replicate the soil moisture-flow relationship in the period of overlap then it is difficult confidently conclude that this relationship is accurately represented prior to the 20th century.

Three questions are addressed by this study: 1) What is the relationship between antecedent fall soil moisture conditions and streamflow in the 100-year-long instrumental record, 2) can antecedent fall soil moisture be skillfully reconstructed utilizing tree rings, and 3) does the soil moisture reconstruction replicate the relationships in the instrumental data in the period of overlap? The relationship between antecedent fall soil moisture conditions and streamflow in the instrumental period were investigated using correlation analysis between gridded November soil moisture and naturalized water-year gaged streamflow at Lees Ferry. We approached the reconstruction using traditional dendrochronological techniques, with water balance model output from the instrumental period as the calibration series.

Using reconstructed autumn soil moisture, we then evaluated whether the reconstruction replicates the antecedent soil moisture and UCRB streamflow relationship present in the instrumental record.

2. Data

2.1. Soil moisture data

Modeled soil moisture data was used for this study because of the limited instrumental coverage for soil moisture in the UCRB and because of the lack of observed records long enough for calibrating a reconstruction model. Initially, we evaluated the output from three hydrologic models that generate soil moisture: the McCabe and Wolock (2011a) water balance model the Fan and van den Dool (2004) Leaky Bucket model and Colorado Basin River Forecast Center model. In all models, the correlation between naturalized water-year Colorado River streamflow at Lees Ferry and monthly modeled soil moisture is weak (˜r < 0.30) in August prior to the onset of the water year but strengthens (˜r > 0.60) in autumn after the water year begins. We selected November as the representative autumn month based on the correlation statistics and because soil moisture is unlikely to vary considerably after the ground freezes and the snowpack begins to accumulate (Huntington and Niswonger, 2012). SNOTEL soil moisture measurements from 13 SNOTEL sites (2003-2012) in the UCRB support this choice (Figure 3), as measured autumn soil moisture generally declines after November and remains low until March (data available from the National Resource Conservation Service at https://www.wcc.nrcs.usda.gov/snow/snotel-data.html). Because all of the soil moisture models demonstrated a strong correlation between antecedent November soil moisture and annual streamflow ($r = >0.51$, $p < 0.01$), and were similar to each other ($r = > 0.80$, $p < 0.01$ for November), we chose to use the longest time series of soil moisture from the McCabe and Wolock water balance model, 1896-2012 (WBM; McCabe and Wolock, 2011a). The WBM is a gridded (4 km x 4 km) hydrologic model that uses an accounting procedure with monthly precipitation and temperature to compute water allocation to components of the hydrologic system including soil-moisture storage, potential and actual evapotranspiration, and snow accumulation and melt (McCabe and Wolock, 2011a). Soil moisture storage output values from the WBM are expressed as available water capacity (cm/cm, cm of water per cm of soil; Dunne and Willmott, 1996). Soil moisture for the grid points in the UCRB has been generated in prior work (Woodhouse et al. 2016). An assessment of the relationship between gridded soil moisture and Colorado River at Lees Ferry water year flow was made. Since relationships are spatially variable, averaging the soil moisture for all UCRB grid cells weakens the November soil moisture to streamflow signal. To reduce the potential for spurious correlations related to the large number of grid cells in the dataset, and to isolate the areas in the basin with the strongest relationship to water-year flow, cells were evaluated based upon a split-period positive and significant correlation threshold ($p<0.01$) with annual Colorado River at Lees Ferry estimated natural flows (U.S. Bureau of Reclamation, https://www.usbr.gov/lc/region/g4000/NaturalFlow/current.html) in both the first half (1906-1958) and second half (1959-2010) of the full period (Figure 4). All grid cells significantly correlated in both halves of the full period were averaged for a UCRB November soil moisture series. The averaged soil moisture series contains first-order AR(1) autocorrelation demonstrating year-to-year persistence in November soil moisture. This series (henceforth referred to as modeled soil moisture) was used to examine relationships between soil moisture and observed streamflow, and for the calibration of the soil moisture reconstruction.

2.2. Streamflow gage data

Naturalized water-year flows (October-September) for the Colorado River at Lees Ferry (1906-2012) were obtained from the U.S. Bureau of Reclamation (https://www.usbr.gov/lc/region/g4000/NaturalFlow/current.html; for more information, see https://www.usbr.gov/lc/region/programs/crbstudy/Report1/TechRptC.pdf). Naturalized flow data estimate flow in the Colorado River with the effects of depletions, diversions, and reservoir management removed. The gage record at Lees Ferry (henceforth referred to as Lees streamflow) is located just above Lake Powell and effectively represents the integrated flow of the UCRB.

2.3. Tree-ring data

We used a network of 62 annual ring-width standard and residual (low order autocorrelation removed) chronologies from trees that are limited in growth by moisture (*Pinus ponderosa*, *Pinus edulis*, and *Psuedotsuga menziesii*) in and adjacent to the UCRB and that were used by Woodhouse et al., (2006) (Figure 4). Both standard and residual tree-ring series were evaluated using correlations with November soil moisture. Standard chronologies were selected based upon higher average correlations with November soil moisture. Standard tree-ring chronologies also contain year-to-year persistence (autocorrelation) and are preferable to residual tree-ring chronologies as reconstruction predictors when the calibration data also contains autocorrelation. All chronologies are publicly available from the International Tree-Ring Data Bank (ITRDB). Correlation analysis was used to screen the pool of 62 standard tree-ring predictors and remove those with statistically insignificant correlations with November soil moisture. From the full pool we selected 26 potential predictors (Table 1) based upon split-period (1906-1958, 1959-2010) positive and significant correlation ($p<0.01$) with November soil moisture. No other screening was undertaken, and consequently, some of the tree-ring chronologies in the pool are the same as those used in the streamflow reconstruction described below.

2.4. Streamflow reconstruction data

Reconstructed streamflow at Lees Ferry, AZ for the period 1490-1997 (Lees-B, Woodhouse et al. 2006) was obtained from the National Oceanic and Atmospheric Administration, Paleo Data Search (available online at https://www1.ncdc.noaa.gov/pub/data/paleo/treering/reconstructions/northamerica/usa/upper-colorado-recons.txt). This reconstruction uses tree-ring chronologies that retain the year to year persistence (standard chronologies) that closely match that of Colorado River streamflow to reconstruct water year streamflow in the UCRB at the Lees Ferry gage.

3. Methods

3.1. Instrumental data analysis

To evaluate the relationship between modeled soil moisture and Lees streamflow series and the climate information in each, we derived correlation coefficients with monthly UCRB precipitation and temperature over the common period (1906-2010). Superposed epoch analysis (SEA) was then used to evaluate the relationship between soil moisture and streamflow during the instrumental period. SEA is a technique employed when many observations of an event of interest exist (e.g. low streamflow) but the noise from other influences may be obscuring any detectible response to that event (Singh, 2006). Compositing of soil moisture conditions at annual lags relative to an anomalous flow in SEA allows the event signal to remain while averaging out other influences, and is used to reveal lagged relationships that are statistically significant. The SEA reveals if soil moisture conditions preceding a high/low flow are drier/wetter than would be expected by chance alone. The *sea* function in *dplR* quantifies the relationship between November soil moisture and streamflow anomalies (Bunn 2008). Event years used in the *sea* function were the years that fell into the wettest (80th percentile) and driest (20th percentile) flow in the Colorado River at Lees Ferry (Table 2). The streamflow event years for the instrumental period (1906-1998; n=18 wet, n=18 dry) were utilized in the SEA with observed soil moisture.

3.2. Reconstruction development and assessment

In order to calibrate a reconstruction model for November soil moisture, we used the pool of 26 candidate tree-ring chronologies and modeled soil moisture in a stepwise multiple linear (least-squares) regression over the years 1906-1998. Model cross-validation was accomplished using a leave-one-out process, withholding one data point from the calibration period and a prediction is made for that point, iteratively for each value in the calibration period (Michaelsen, 1987). The validation statistics Reduction of Error (RE) and Root Mean Square Error (RMSEv) describe the skill of the model (Fritts, 1976; Cook et al., 1999). The reduction of error (RE) statistic is a measure of reconstruction reliability used during model verification (Cook and Kairukstis, 1990). The explained variance statistic in the calibration process is similar to the RE used during validation, which provides an average estimation to be compared to the regression estimate. RE values greater than zero indicate skill in the reconstruction relative to a nave estimate of the long-term mean (Fritts, 1976). Using the sign test (Dixon and Mood, 1946) we tested the reconstruction estimates for the frequency of agreement between the signs of the departures of the estimated and observed data from the sample mean. Significance of the sign test ($p < 0.01$) is determined following the cumulative distribution tables for the binomial distribution, indicating that positive signs occur more frequently than would be expected by chance alone (Beyer, 1968).

3.3. Tree-ring based streamflow-soil moisture inflation

To understand the influence of November soil moisture on streamflow variability over the reconstruction period, our goal was to produce a soil moisture reconstruction that could be compared with an existing streamflow reconstruction. However, the November soil moisture reconstruction and previously developed Lees Ferry streamflow reconstruction (LeesB; Woodhouse et al. 2006) share one of the same chronology predictors (NPU). Because of this, the relationship during the common period (1906-1998) between reconstructed soil moisture and reconstructed flow is inflated (observed $r = 0.51$, reconstructed $r = 0.83$) which may influence results in further analyses. To determine if the reconstruction is conveying the same underlying climate information as the instrumental record and to gain a more complete understanding of the shared information between soil moisture and streamflow, we assumed that the shared information is climate (precipitation and temperature). We then evaluated the November soil moisture record statistically independent of streamflow for both the observed and reconstructed series. To do this, the shared variance between soil moisture and streamflow was removed using linear regression. To quantify the effects of removing this statistical dependence and facilitate interpretation of the soil moisture-flow relationship, observed and reconstructed values (pre-regression) and their residuals (post-regression) were correlated with monthly instrumental UCRB precipitation and temperature in the antecedent months (prior July through prior November), over the common period (1906-1998). Climate information remaining in the soil moisture residuals is climate information influencing soil moisture that is not shared with flow. The difference between statistically significant ($p < 0.05$) climate correlations before regression and those remaining after regression allows us to approximate the monthly climate information shared between soil moisture and flow in both the observed and reconstructed series.

3.4. Comparison of the reconstructed soil moisture and streamflow relationship: SEA

To evaluate if reconstructed November soil moisture replicates the soil moisture-flow relationship in the instrumental record, we performed SEA analysis on the common period of reconstructed November soil moisture. Event years from the Lees streamflow reconstruction series (1906-1998; n=18 wet, n=18 dry) were utilized in the SEA with the soil moisture reconstruction series. Event years were the years that fell into the wettest (80th percentile) and driest (20th percentile) flow in the Colorado River LeesB reconstruction (Table 2). Soil moisture values for the five years preceding each event year were averaged. The average for each year lag was plotted, and significance levels were determined using bootstrap resampling (n=1000).

3.5. Comparison of the soil moisture and streamflow relationship: Counts

The SEA analysis shows statistically significant average conditions at annual time lags from an anomalous flow, but it can provide misleading results if statistical artifacts affect the composite calculations. We used counts analysis as a non-parametric approach to evaluating the frequency of soil moisture occurrence up to four years from a low flow or three years from a high flow and to compare against SEA results. Counts analysis is different than SEA because counts examine the *frequency* of soil moisture conditions versus SEA which is *averaged* conditions at specified lags without consideration of how often individual dry or wet soil moisture years occur at each lag, or if conditions persist across lags. Our counts analysis examined the frequency of soil moistness categories in years prior to a low/high streamflow year (hereafter referred to as same/t-0, t-1, t-2, t-3 and t-4) in both the observed and reconstructed records. November soil moisture was categorized by quintile (Very dry: 0-20%, Dry: 21-40%, Moderate: 41-60%, Wet: 61-80%, Very wet: 81st-100%). The years in the bottom quintile of flow and in the top quintile of flow were used as the low flow and high flow years, respectively. The streamflow years in the bottom quintile and top quintile were then compared to counts in soil moisture categories in the same, t-1, t-2, t-3, and t-4 years. The occurrence of categorized observed and reconstructed soil moisture years in each streamflow quintile division was counted for the common period (1906-1998).

Given that the most plausible relationship between antecedent soil moisture and anomalous flow is in the same water year, contingency tables and the Pearsons Chi-Square test were used to determine if there is a significant ($p < 0.05$) relationship in counts between soil moisture categories and low flow or high flow in the same water year. The Chi-Square test evaluates tests of independence, which compare the pattern of counts in the data to the pattern that would be expected if the counts were independent. The Chi-Square statistic is the square of the difference between what is observed in the data and what would be expected divided by the expected value. The Chi-Square statistic is calculated and compared with a critical value from the Chi-Square distribution. The null hypothesis of the Chi-Square test is that no relationship exists between soil moisture and streamflow anomalies. A statistically significant result indicates that there is a statistically significant relationship between soil moisture conditions and anomalous flow.

4. Results and Discussion

4.1. Instrumental analysis

Figure 3 shows a time series of instrumental soil moisture data for the UCRB. The monthly time series of modeled soil moisture output in four UCRB sub-basins, modeled soil moisture output in the entire UCRB, and measured SNOTEL soil moisture averaged over 13 sites shows that soil moisture values in the upper snow-dominated sub-basins (e.g. Green, SNOTEL) stop increasing from summertime lows in November, and remain low until February, with the exception of the San Juan sub-basin. Soil moisture in the entire UCRB increases slightly between November and February, but the increase is slow relative to the rate after February. November soil moisture, therefore, captures autumn conditions assuming the ground freezes and stays frozen until spring melt.

November soil moisture grid cells that have stable split-period significant correlation with annual streamflow over the last century are centrally located in the basin (Figure 4). Averaged November modeled soil moisture grid cells and Colorado River water-year flow (Lees streamflow) are significantly correlated ($r = 0.513$, $p < 0.01$). These results are consistent with Tang and Piechota (2009) annual soil moisture in the UCRB using 3 layers of 50 year-long gridded soil moisture obtained from the Variable Infiltration Capacity (VIC) model (Tang and Piechota, 2009; Fig. 8). The central basin area that we identified also exhibits lower soil moisture anomalies during drought years in the Tang and Piechota (2009) soil moisture anomaly maps as well as a sensitivity to wet conditions (Tang and Piechota, 2009). The region of our highest correlation between observed November soil moisture and gaged streamflow also corresponds to Anderson et al. (2012) Region 3 regionalized annual soil moisture from the Fan and Van den Dool (2004) model. Andersons Region 3 produced, in part, the highest amount of explained variance in their tree-ring reconstructions of annual soil moisture compared with all four regions they examined. These results support our split-period correlation, and correspond to the location of our reconstruction predictors and their explained variance.

SEA results for the observed record show a significant statistical relationship between dry Nov soil moisture and low Lees streamflow (20th percentile) event years (Figure 5a and 5b; Table 5) in the same year (0; $p < 0.01$) and one year before the low flow year (-1; $p < 0.05$). Antecedent soil moisture the year of and one year preceding a low Lees streamflow year is drier than would be expected by chance alone, and driest in the same year as a low flow. In high flow event years (80th percentile), a significant statistical relationship ($p < 0.01$) is present in wet November soil moisture in the same year (0) and one year prior (-1) to a high flow. Soil moisture conditions are on average wettest in the same year as a high flow. Although the SEA composite analysis might imply a direct relationship between November soil moisture and streamflow years later, these results are not indicating a higher frequency of soil moisture wetness years in advance of a flow. Instead, it shows if averaged soil moisture at each lag is significantly dry/wet before a low/high flow.

4.2. Reconstruction development and assessment

The calibration period for the November soil moisture reconstruction for the UCRB using ring-width tree-ring chronologies is 1906-1999. Predictors in the final reconstruction model are the COL, NPU, DOU and RED chronologies (descriptions of the chronologies in the pool of predictors are listed in Table 1). The residuals from the reconstruction are normally distributed. The Durbin-Watson (D-W) test for autocorrelation in the regression residuals is 1.38, indicating that we must reject the null hypothesis that there is zero first order autocorrelation in the residuals ($p < 0.01$, one-sided). The root mean square error of validation (RMSEv) is 14.2973 cm/cm (available water capacity). An R^2 value that is similar to an RE value indicates model robustness. Our November soil moisture reconstruction model explains 51% of the variance ($R^2 = 0.51$) in the calibration period, and the RE is 0.46 (Table 3a). Sign test results indicate that the sign of the estimate is more often correct than would be expected from chance alone (critical value $p = 0.99$; 56 positive/35 negative). Comparison statistics between the instrumental record and the reconstruction show a narrowed range of variability in the reconstruction versus the instrumental record (Table 3b; Max, Min, SD). Soil moisture in the instrumental period, as well as the reconstructed common period and the full reconstruction, show significant first order autocorrelation (AC(lag 1)). The full reconstruction spans 513 years, 1486-1999 (Figure 6).

4.3. Consideration of the streamflow-soil moisture relationship

We found that soil moisture has a distinct relationship to streamflow with unique similarities and differences between the observed and reconstructed series, and soil moisture and streamflow share components of the hydroclimatic system that similarly contribute to tree growth (Fritts, 1976). Table 4 shows correlations of modeled

November soil moisture, reconstructed soil moisture (NovSM), observed streamflow (Lees streamflow) and reconstructed streamflow (LeesB) and their residuals. Instrumental and reconstructed soil moisture are highly and significantly correlated ($r = 0.714$, $p < 0.01$), as are instrumental and reconstructed streamflow ($r = 0.915$, $p < 0.01$). The boxes in Table 4 highlight the soil moisture-flow correlations in the instrumental and reconstruction records. When the influence of soil moisture on flow is removed through regression, correlations between soil moisture residuals and streamflow are weak and no longer significant.

Correlations of the observed soil moisture with monthly precipitation show that soil moisture and streamflow share about 23% of the precipitation information from the prior summer (July) (Figure 7). The observed records also share about 1/3 of the precipitation information in October and November. Correlations of the observed series and residuals also show that a large proportion (~87%) of temperature information is shared in prior July. Autumn temperature (Sep-Nov) is also shared between the instrumental series, but to a lesser amount than in summer (~30%). Reconstructed soil moisture shares a high proportion of autumn (Sept-Nov) precipitation information (~49-84% of the correlation) with LeesB flow. About 95% of the temperature information shared between reconstructed soil moisture and LeesB flow is shared in July, and about 58% of the temperature information is shared in September.

Our results demonstrate that observed Colorado River streamflow and modeled soil moisture have a similar relationship to monthly precipitation (autumn) and monthly temperature (mid-late summer). The shared correlation between modeled soil moisture and gaged flow highlight this shared climatic influence; with a significant portion of the shared correlations corresponding to relationships between soil moisture and climate and flow and climate independently. Our results also reveal differences between the observed and reconstructed records. Tree-ring reconstructions may overestimate extreme reconstructed values, especially in dry years. The tree-ring-based serial correlation also means that the relationships seem stronger in the reconstruction than in the instrumental correlations. These differences complicate our ability to make definitive conclusions about the influence of November soil moisture conditions on low or high flows.

4.4. Comparison of the reconstructed soil moisture and streamflow relationship: SEA

SEA results for the common period of the reconstruction (1906-1998) show a statistical relationship between November soil moisture and reconstructed Colorado River streamflow at certain lags (Figure 5c and 5d; Table 5). Average antecedent soil moisture two years prior to low flow in the Colorado River is significantly drier than would be expected by chance alone ($p < 0.01$). All years up to lag t-3 are on average wet enough prior to a high flow to be significantly different from random variability ($p < 0.05$), with a higher significance ($p < 0.01$) in one, two, and three years.

These results do not entirely replicate the SEA results from the instrumental record but they do suggest antecedent soil moisture is likely influencing flow because of the statistical pattern of drier or wetter soil moisture in years before an anomalous flow in all results. The instrumental record indicates that soil moisture conditions in the same year show the greatest deficit/surplus while the reconstructed common period indicates that soil moisture conditions at lag t-2 show the greatest deficit/surplus.

4.5. Comparison of the soil moisture and streamflow relationship: Counts

Counts of antecedent soil moisture in the instrumental record indicate that very dry conditions (0-20th pctl) occur most frequently in t-1 and t-3 years as a low flow (6 counts; Figure 8). Dry conditions (21-40th pctl) occur most frequently in the same year as a low flow (8 counts). All other counts categories in the same year as a low flow are half of or fewer than the dry counts relative to a low flow year. Counts in t-1 years relative to a low flow vary among all antecedent soil moisture conditions (3-6 counts) with peaks in very dry and dry conditions, with zero occurrences in very wet conditions. In t-2 years prior to a low flow, results are similar including fewer counts in the wet and very wet categories and highest occurrence in dry and moderate conditions. In t-3 years relative to a low flow, soil moisture is most frequently very dry (6 counts). In t-4 years prior to a low flow, counts of antecedent soil moisture are within one or two occurrences, with the exception of wet years (1 count). Counts of soil moisture in each category for years prior to a high flow year in the observed record indicate that wet and very wet soil moisture conditions are most common at all lags (Figure 8). In all previous years relative to a high flow, with the exception of dry conditions in t-2 and t-3 years, wet antecedent soil moisture counts are greater than 3, indicating that antecedent soil moisture relative to a high flow in the observed record is most commonly wet.

Counts of antecedent soil moisture in each category for years prior to a low Colorado River flow year during the reconstructed common period indicate that very dry soil moisture occurs most frequently in the same year as a low flow (Figure 8). Apart from the same-year counts with very dry conditions, all other quintile conditions at all other lags are comparable (12-25 counts). Very dry conditions at lag t-1 are slightly higher than the other counts (34 counts) and very wet conditions are the lowest at all lags. Counts of antecedent soil moisture in the same year as a high flow are also dominated by the very wet (80th percentile) category (Figure 8). Similar to low flow results, the opposing soil moisture conditions (e.g. dry soil moisture prior to high flow years) are at or near zero in the same year. Counts of antecedent soil moisture conditions one year prior to a high flow is most frequently very wet as well (34 counts). Counts of antecedent soil moisture two to three years prior to a high flow vary between 26 counts (moderate) and 10 counts (very dry).

In low flow years, counts of moderate soil moisture conditions two years prior occur with the same frequency as dry conditions (7 counts). Though moderate soil moisture counts in the same year as a low flow are fewer than dry soil moisture, they are four times as frequent as a wet year. The occurrence of moderate soil moisture conditions prior to a high flow in the instrumental record is not as frequent. Counts of instrumental antecedent soil moisture in the middle percentiles (dry-moderate-wet) show more variability than average or near average conditions in the reconstructed series. This suggests that the instrumental record is capturing moderate soil moisture conditions and their relationship to streamflow, especially in years relative to a low flow, more effectively than the tree-ring based reconstruction.

The proportions of t-x counts in wet years (wet + very wet) relative to high flow, and in dry years (dry + very dry) relative to low flow, suggest a strong relationship between anomalous flow years and soil moisture conditions in the same year (Table 6). These proportions exceed 70% in all categories, and in both the long-term reconstruction and in the common period (1906-1998). Observed series (modeled soil moisture and Lees streamflow) show a higher proportion of dry and wet soil moisture conditions in t-1 year relative to a low or high flow, respectively, than proportions over the same period in the reconstruction. By contrast, the proportion of dry soil moisture occurrences two to three years prior to a low flow in the observed record is smaller than the reconstructed occurrences in the common period.

Given the plausible relationship between antecedent soil moisture and anomalous flow in the same water year, and based upon results of the SEA and counts, we used contingency tables and Chi-Square to test antecedent soil moisture and anomalous flow in the same water year. The Chi-Square null hypothesis states that no relationship exists between soil moisture and streamflow anomalies. Results indicate that we must reject the null hypothesis for the observed and reconstructed counts (Table 7). There is a statistically significant ($p < 0.05$) relationship between soil moisture categories and streamflow anomalies in both the observed and reconstructed records. The Pearson Chi-Square statistic for the observed counts is 24.683 (df =

4), and Pearson Chi-Square statistic for the reconstructed counts is 33.569 (df = 4). The largely disproportionate association of dry soil moisture to low flow, or wet soil moisture to high flow, is indicated in Table 7 as percent count within soil moisture (SM). While these results show an association between soil moisture conditions and streamflow anomalies in both the observed and reconstructed counts, the observed contingency table demonstrates associations that are distributed across the counts (e.g. very dry and moderate-wet soil moisture counts are associated with both low and high flows) with the exception of dry soil moisture counts and high streamflow or very wet soil moisture counts and low streamflow. By contrast, the reconstructed counts associated with low streamflow are limited to only very dry and dry soil moisture. Similarly, for high streamflow, all of the reconstructed counts are associated with moderate, wet and very wet soil moisture. These results accentuate the tree-rings potential inability to replicate average or near-average conditions found in the instrumental record.

There is likely not a relationship between antecedent soil moisture conditions in years preceding flow without the influence of intervening years. SEA (intensity) and counts (frequency) analyses do not adequately resolve runs of soil moisture conditions, or the sequencing of soil moisture from one year to the next. We examined intervening years in our results and found that in the reconstructed common period (1906-1998) most intervening years are in the same percentile category as both the lagged year and the flow year. At larger lags, the probability of opposite soil moisture conditions (wet to dry/dry to wet) occurring between the lagged year and the flow year increases. The low flow year of 1965 is an example of an anomalous low flow year that is preceded by mixed soil moisture years. Two years prior to the low flow, soil moisture is in the 20th percentile. However, the intervening year is in the 80th percentile, and then soil moisture in the flow year returns to the 20th percentile. The low flow year of 1968 is a second example. Two years prior to the low flow, soil moisture is in the 20th percentile. However, the intervening year and the flow year are in the 60th percentile. Different combinations of influences on streamflow (e.g. spring temperature, cool-season precipitation, base flow) not including November soil moisture could be moderating total water year flow (Woodhouse et al. 2016). These preliminary findings suggest that while antecedent conditions occurring prior to the anomalous flow year are likely to be consistent in wetness with the anomalous flow, instrumental SEA analysis does not convey a lagged relationship more than one year prior to an anomalous flow. Persistence in the tree-ring data may be extending the lagged relationship in the reconstructed analysis. Therefore, a direct causal link between lagged antecedent soil moisture cannot be determined here and further research is required.

4.6. Does the soil moisture reconstruction replicate the relationships in the instrumental data?

A tree-ring based reconstruction was generated for this study to demonstrate the potential for reconstructing antecedent soil moisture, and to assist with evaluating if reconstructions can replicate instrumental hydroclimatic relationships in the UCRB. Our results suggest that a hydrologic relationship between antecedent soil moisture and water-year flow is present in the instrumental data. While our analysis also suggests a statistically significant relationship between soil moisture and streamflow extremes in the reconstructed series, the analyses used in this study do not produce results that closely replicate patterns in the relationship exhibited the instrumental record. One known characteristic of tree-rings is the ability to capture extremes, especially dry extremes. When evaluating the full range of soil moisture conditions found in instrumental record, tree-rings fail to capture and effectively replicate conditions closer to average.

Inconsistency in results between the observed records and reconstructed records can be explored in the uncertainties of the data used to generate the reconstruction. Patterns in soil moisture are difficult to analyze because soil moisture integrates the inputs from precipitation and the losses from evapotranspiration, surface runoff, and baseflow (Hamlet et al., 2007). Soil moisture is also characterized by autocorrelation, which means that lagged effects in inputs or losses may be equally as important as those occurring at the time of observation (Clyde, 1940). We used soil moisture data generated from a hydrologic model with inherent sources of uncertainty. Sources of uncertainty in this model may contribute to the differences observed in our results, including the presence of any errors in the precipitation and temperature input data, as well as the potential influence of land use and water management on estimated runoff (McCabe and Wolock, 2011b). Naturalized flow also introduces potential error in our results because flow naturalization makes estimations to account for consumptive uses and losses that affect aggregate flow in the basin including agricultural uses and reservoir regulation (USBR, 2012).

An inflated association between the two reconstructions themselves may drive differences between the observed record and the reconstructed results as well. Tree-ring based reconstructions have inherent errors that arise during the chronology development process and as a result of multiple linear regression modeling as stated above (Beyer, 1968; Fritts, 1976; Michaelsen, 1987; Cook and Kairukstis, 1990; Cook et al., 1999). As environmental proxies, trees are influenced by ecological as well as climatic factors (Fritts, 1976). Several statistical assumptions are made for regression modeling, which were stated and met for this study. However, for both reconstructions utilized here, each respective regression model explains only a portion of the total variance in the observed series. Uncertainty within the unexplained variance, compounded by the comparison of two reconstructions, may contribute to differences observed in our results.

5. Conclusion

There is a clear, statistical relationship between autumn soil moisture and UCRB water-year flow. A 513-year long November soil moisture reconstruction demonstrates that antecedent soil moisture can be successfully reconstructed using moisture-sensitive tree-rings in the UCRB. Our results show a robust reconstruction model explaining 51% of the variance in November soil moisture. The November soil moisture reconstruction shows some dependence on the streamflow reconstruction utilized in our analysis because of the shared tree-ring derived variance. By removing this dependence and analyzing the residuals we established the shared climate relationship between soil moisture and streamflow. Instrumental soil moisture and streamflow share climate information that is replicated, but also enhanced, in the reconstructions. Instrumental soil moisture conditions show a significant association up to one year prior to an anomalous flow on the Colorado River. It is not clear from all SEA and counts analyses what lag may be most important for estimating a high or low flow year, but these results suggests a significant relationship between antecedent soil moisture and streamflow in both the instrumental and reconstructed records. The reconstruction of antecedent soil moisture reveals differences in the soil moisture-streamflow relationship between the instrumental record and the tree-ring based reconstruction. Further research is essential to untangle the influences of antecedent soil moisture on streamflow during the past 100 years to aid in interpretation of long-term records. In the future, reconstructions of antecedent soil moisture may inform monitoring efforts and forecasts of flow anomalies on the Colorado River when soil moisture deficit or surplus is observed in the same year, or one and two years in advance. These long-term records will also contribute to our understanding of hydrologic relationships in large river basins in the past and in the future.

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Tables

Table 1 Potential tree-ring predictor chronologies used in the stepwise linear regression for the November soil moisture reconstruction.

Code	Name	Species	Begin	End	Lat (deg)	Lat (min)	Long (deg)	Long (min)	Elev (m)
COD	Cochetopa Dome	Ponderosa pine	1437	2002	38	15	106	40	2835
SAP	Sapinero Mesa	Ponderosa pine	1511	2000	38	19	107	12	2700
BRR	Boulder Ridge Rd	Ponderosa pine	1423	2001	40	59	105	40	2650
SFK	South Fork	Ponderosa pine	1566	2002	37	40	106	39	2591
SPP	Soap Creek	Ponderosa pine	1541	1999	38	32	107	19	2417
SPC	Sheep Pen Canyon	Ponderosa pine	1460	1998	37	4	103	16	1585
WIL	Wild Rose	Pinyon pine	1146	2002	39	1	108	14	2636
RPC	Red Pine Canyon	Pinyon pine	1405	2001	40	37	109	57	2335
NUR	Nutter's Ridge	Pinyon pine	1040	2001	39	49	110	40	2235
UNA	Unaweep Canyon	Pinyon pine	1296	2000	38	50	108	34	2225
TRG	Trail Gulch	Pinyon pine	1402	2002	39	43	106	59	2210
MCP	McPhee-Dolores	Pinyon pine	1270	2002	37	35	108	35	2195
PUM	Pump House	Pinyon pine	1320	2002	39	58	106	31	2194
DJM	Dutch John Mtn.	Pinyon pine	1365	2001	40	58	109	25	2190
RED	Red Canyon	Pinyon pine	1336	1999	39	42	106	44	2164
WED	Well's Draw	Pinyon pine	887	2001	39	50	110	10	2145
PLU	Plug Hat Butte	Pinyon pine	1270	2000	40	47	108	58	2133
RIF	Rifle	Pinyon pine	1335	2000	39	40	107	53	2073
COL	Collins Gulch	Pinyon pine	1168	2001	39	50	108	12	2050
SLK	Slickrock	Pinyon pine	1490	2002	38	1	108	55	2000
DRY	Dry Park	Pinyon pine	1536	2000	38	15	108	20	1996
PIC	Piceance	Pinyon pine	1124	2001	40	3	108	18	1900
EFU	Escalante Forks update	Pinyon pine	1569	1999	38	39	108	20	1737
DOU	Douglas Pass	Douglas-fir	1382	2000	39	36	108	48	2591
ENC	Encampment	Douglas-fir	1380	2001	41	9	106	47	2500
NPU	North Park update	Douglas-fir	1486	2001	40	57	106	20	2450

Table 2 High Colorado River streamflow (80th pctl) and low Colorado River streamflow (20th pctl) event years used for the SEA instrumental period, 1906-1998 (columns 1 and 2, respectively) and for the SEA reconstructed common period, 1906-1998 (columns 3-4, respectively).

Instrumental		Reconstructed (common)	
80pctl Common	20pctl Common	80pctl Common	20pctl Common
1907	1931	1906	1908
1909	1934	1907	1934
1914	1940	1909	1936
1916	1946	1914	1940
1917	1954	1917	1946
1920	1955	1920	1953
1921	1959	1921	1954
1929	1961	1929	1956
1942	1963	1942	1959
1952	1964	1949	1961
1957	1966	1952	1963
1973	1976	1957	1964
1983	1977	1973	1966
1984	1981	1983	1968
1985	1989	1984	1976
1986	1990	1985	1977
1995	1992	1986	1981
1997	1994	1987	1990

Table 3 Calibration, validation, and comparison statistics for the November soil moisture reconstruction (1486-1998). a) Calibration statistics are the coefficient of determination (R^2), the R^2 adjusted for the number of predictors in the model (R^2a) and Standard Error of the Estimate (SEE). Validation statistics are Reduction of Error (RE) and Root Mean Square Error of the validation (RMSEv). b) Comparative descriptive statistics of soil moisture in the instrumental period (1906-1998), the common reconstruction period (1906-1998), and the full reconstruction (1486-1998). Further description of reconstruction statistics in the text.

a)

Calibration statistics			Validation statistics				
R^2	R^2_a	SEE	RE	RMSE	Sign Test (hit/miss)	Predictors	Std. Coefficients
0.514	0.492	13.9155	0.46	14.3	56/35*	COL	0.287
						NPU	0.193
						DOU	0.232
*Significant at p<0.01						RED	0.218

b)

	Mean	Max	Min	SD	AC(lag 1)	AC signif.
Instrumental	41.076	95.658	10.07	19.417	0.519	0.000
Recon Common	41.076	81.655	8.594	13.921	0.523	0.000
Recon Full	38.185	81.655	-5.15	15.392	0.447	0.000

Table 4 Correlations of observed soil moisture with observed streamflow (1906-2010) and reconstructed soil moisture with reconstructed streamflow (1906-1998) and with the corresponding observed/reconstructed soil moisture residuals (Resid). The two boxes highlight correlations between instrumental (Obs) and reconstructed (Recon) records.

	Soil Moisture Obs	Soil Moisture Recon	Flow Obs	Flow Recon	Soil Moiture Obs Resid
Soil Moisture Recon	.714**				
Flow Obs	.513**	.737**			
Flow Recon	.584**	.830**	.915**		
Soil Moiture Obs Resid	.858**	.404**	0	0.146	
Soil Moisture Recon Resid	.430**	.578**	-0.015	0.024	.517**

** Correlation is significant at the 0.01 level (2-tailed).

Table 5 Superposed Epoch Analysis (SEA) results for the reconstructed series. The lag column indicates the negative lag year from the event year. The se column indicates the average composite value for soil moisture residuals at that lag, and se.unscaled is the same values without standardization. The p column indicates the significance level (p <) determined using bootstrap resampling (n=1000) at each lag.

20th Percentile (1906-1998)				**20th Percentile (1490-1998)**			
lag	**se**	**se.unscaled**	**p**	**lag**	**se**	**se.unscaled**	**p**
-5	-0.149	35.889	0.279	-5	-0.195	35.175	0.04
-4	-0.230	34.644	0.149	-4	-0.039	37.580	0.363
-3	-0.128	36.209	0.316	-3	-0.123	36.293	0.127
-2	-0.884	24.559	0	-2	-0.575	29.330	0
-1	-0.083	36.913	0.385	-1	-0.872	24.749	0
0	-0.068	37.137	0.429	0	-0.452	31.228	0

80th Percentile (1906-1998)				**80th Percentile (1490-1998)**			
lag	**se**	**se.unscaled**	**p**	**lag**	**se**	**se.unscaled**	**p**
-5	0.395	44.274	0.045	-5	0.001	38.195	0.475
-4	0.511	46.055	0.012	-4	0.117	39.984	0.083
-3	0.619	47.721	0.009	-3	0.355	43.659	0
-2	1.112	55.319	0	-2	0.638	48.012	0
-1	0.745	49.662	0.002	-1	0.976	53.228	0
0	0.436	44.897	0.023	0	0.356	43.673	0

Table 6 Percentage of counts at dry (0-40%) and wet (61-100%) categorical soil moisture lags relative to year of low/high flow, respectively. The Dry + Very Dry columns are proportion of counts relative only to low flow years. The Wet + Very Wet columns are proportion of counts relative only to high flow years.

	**(1906-1998) Dry + Very Dry	**(1906-1998) Wet + Very Wet	*(1906-1998) Dry + Very Dry	*(1906-1998) Wet + Very Wet	*(1490-1998) Dry + Very Dry	*(1490-1998) Wet + Very Wet
Same Flow Year (t)	72%	83%	100%	83%	95%	92%
t-1 Year	61%	78%	56%	50%	54%	56%
t-2 Year	39%	47%	56%	50%	41%	47%
t-3 Year	39%	53%	56%	56%	42%	43%
t-4 Year	44%		45%		55%	

* Reconstructed November Soil Moisture Residuals and LeesB Reconstruction
** McCabe November Soil Moisture Residuals and Gage Record

Table 7 Contingency table and Pearson Chi-Square statistics for counts during the common period at all categorical soil moisture conditions in the a) observed (1906-1998), and b) reconstructed (1906-1998) series relative to year of very dry (low)/very wet (high) flow. Significance level $p < 0.05$.

a)

Obs Soil Moisture		**Obs Flow Anom**		
		Low	High	Total
Very dry	Count	8	1	9
	% within SM	88.9%	11.1%	100.0%
Dry	Count	5	0	5
	% within SM	100.0%	0.0%	100.0%
Moderate	Count	4	2	6
	% within SM	66.7%	33.3%	100.0%
Wet	Count	1	6	7
	% within SM	14.3%	85.7%	100.0%
Very wet	Count	0	9	9
	% within SM	0.0%	100.0%	100.0%
Total	Count	18	18	36
	% within SM	50.0%	50.0%	100.0%

Pearson Chi-Square	**df**	**Significance**
23.683	4	p = 0.000

b)

Recon Soil Moisture		**Recon Flow Anom**		
		Low	High	Total
Very dry	Count	13	0	13
	% within SM	100.0%	0.0%	100.0%
Dry	Count	6	1	7
	% within SM	85.7%	14.3%	100.0%
Moderate	Count	0	2	2
	% within SM	0.0%	100.0%	100.0%
Wet	Count	0	5	5
	% within SM	0.0%	100.0%	100.0%
Very wet	Count	0	10	10
	% within SM	0.0%	100.0%	100.0%
Total	Count	19	18	37
	% within SM	51.4%	48.6%	100.0%

Pearson Chi-Square	**df**	**Significance**
33.569	4	p = 0.000

Figures

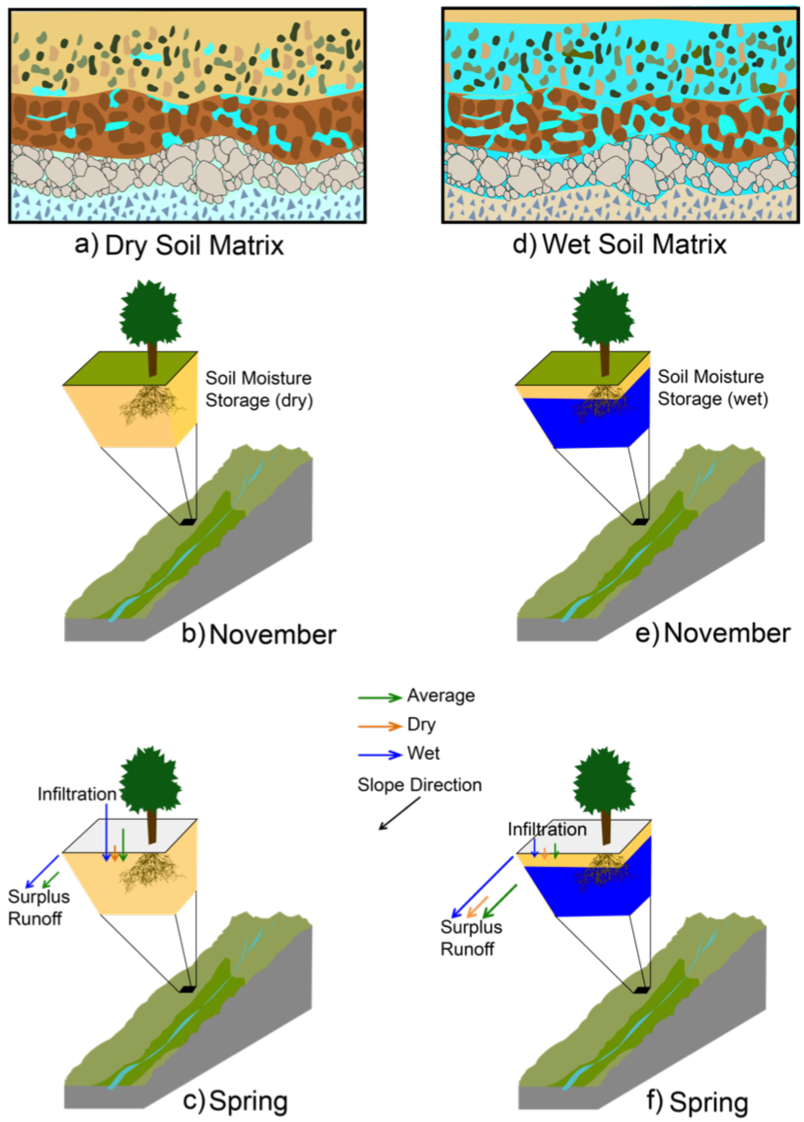

Figure 1 *November Soil Moisture Conceptual Diagram.* Soil moisture storage capacity is the total amount of potential storage in the WBM one-meter profile and is indicated in brown. Ice in the soil matrix is pale blue: a) lesser ice congestion and d) more ice-congestion. a-b) Dry soil moisture when the soil freezes in November, c) relative contributions of spring snowmelt to infiltration and runoff. The length of each arrow represents volume of soil moisture recharge versus surplus runoff in the spring (more/longer, less/shorter) and arrow color is the relative contribution during average (green), dry (brown) and wet (blue) winter precipitation conditions. Water in soil moisture storage is indicated in dark blue. d-f) as in (a-c) but with wet soil moisture when the soil freezes in November.

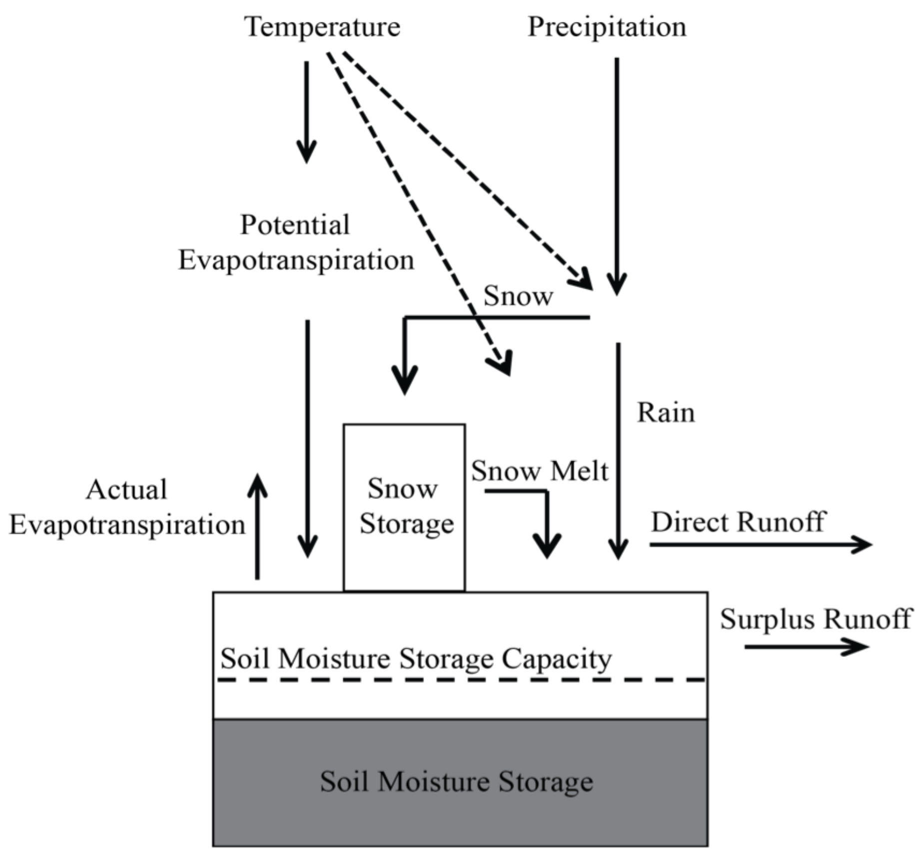

Figure 2 Diagram of the McCabe and Wolock hydrologic model, including soil moisture as one product of the model. (McCabe and Wolock, 2011a)

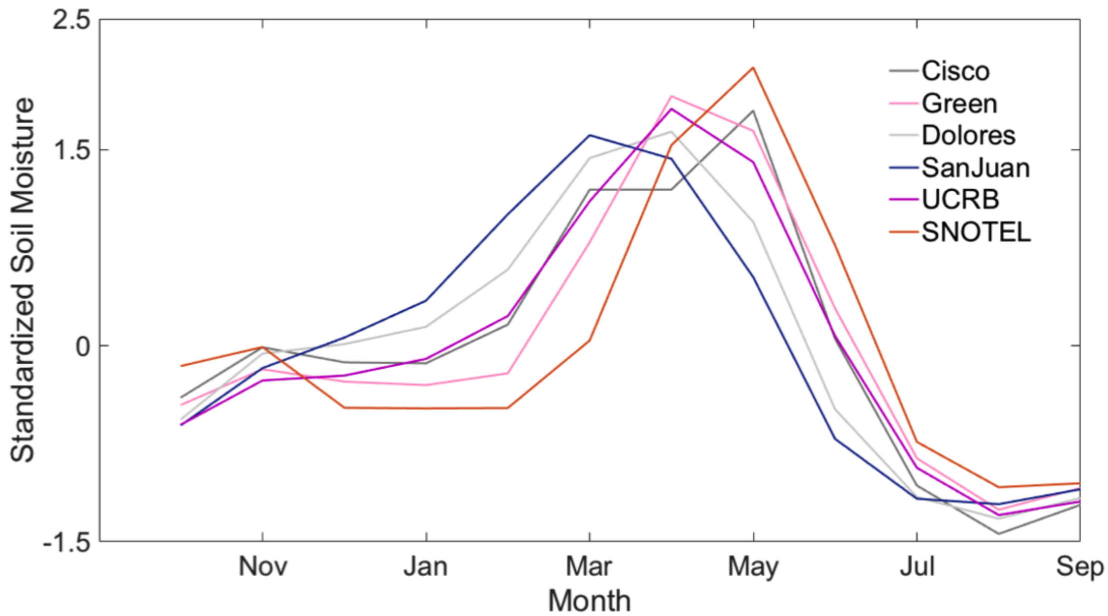

Figure 3 Monthly soil moisture from the hydrologic model (McCabe and Wolock, 2011) in four UCRB sub-basins and the entire UCRB, and measured soil moisture at 13 SNOTEL sites (2003-2012).

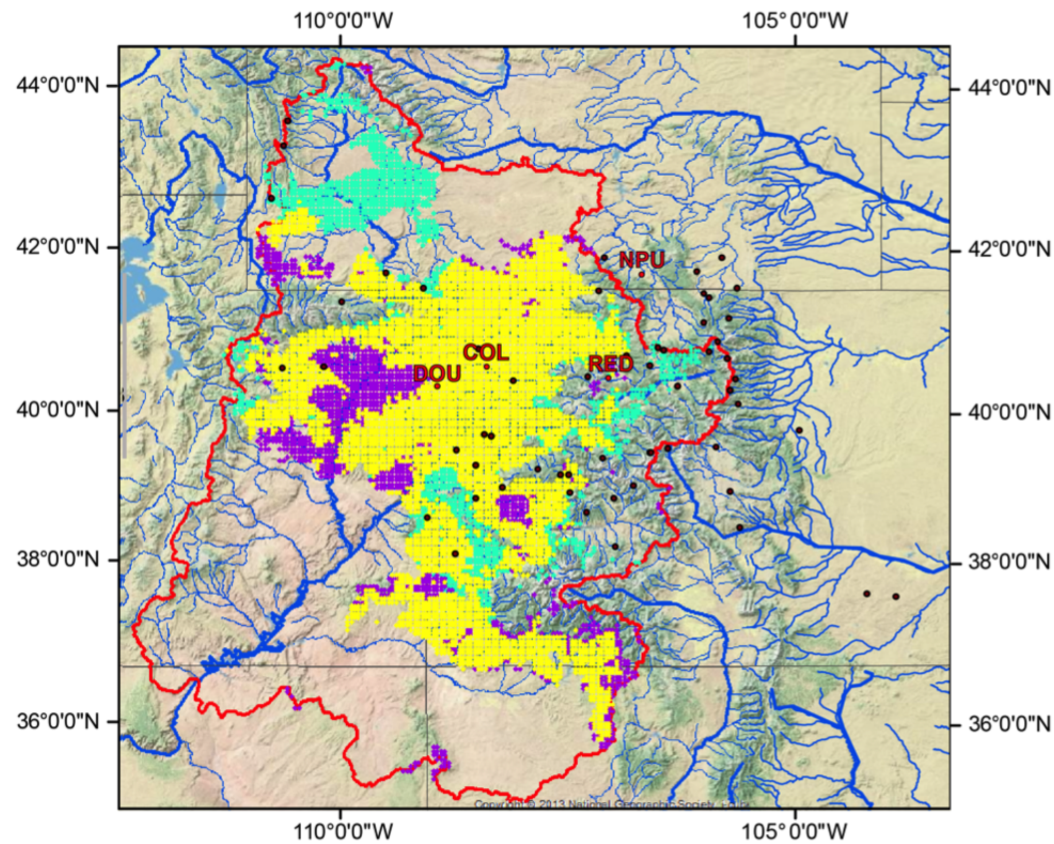

Figure 4 Significant (p<0.01) correlations between November soil moisture grid cells and Lees Ferry flow in the first half of the period (purple cells, 1906-1958), in the second half of the period (green cells, 1959-2010) and in both (yellow cells). Dark red dots are all tree-ring chronologies considered in this study. Red dots with red labels are the four predictor chronologies used in the reconstruction model.

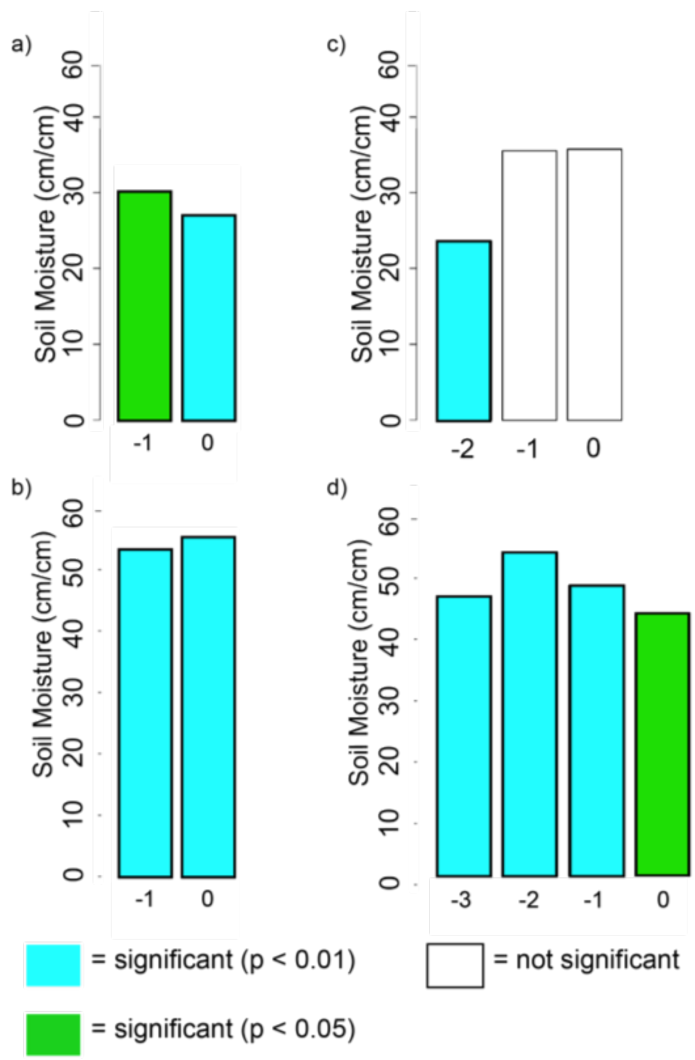

Figure 5 Superposed Epoch Analysis (SEA) for observed November soil moisture (1906-1998) and gaged streamflow at Lees Ferry (a-b), and reconstructed November soil moisture (common period 1906-1998) and reconstructed Lees streamflow (c-d). The 20th percentile (dry) event years are shown in panels a) and c). The 80th percentile (wet) event years are shown in panels b) and d).

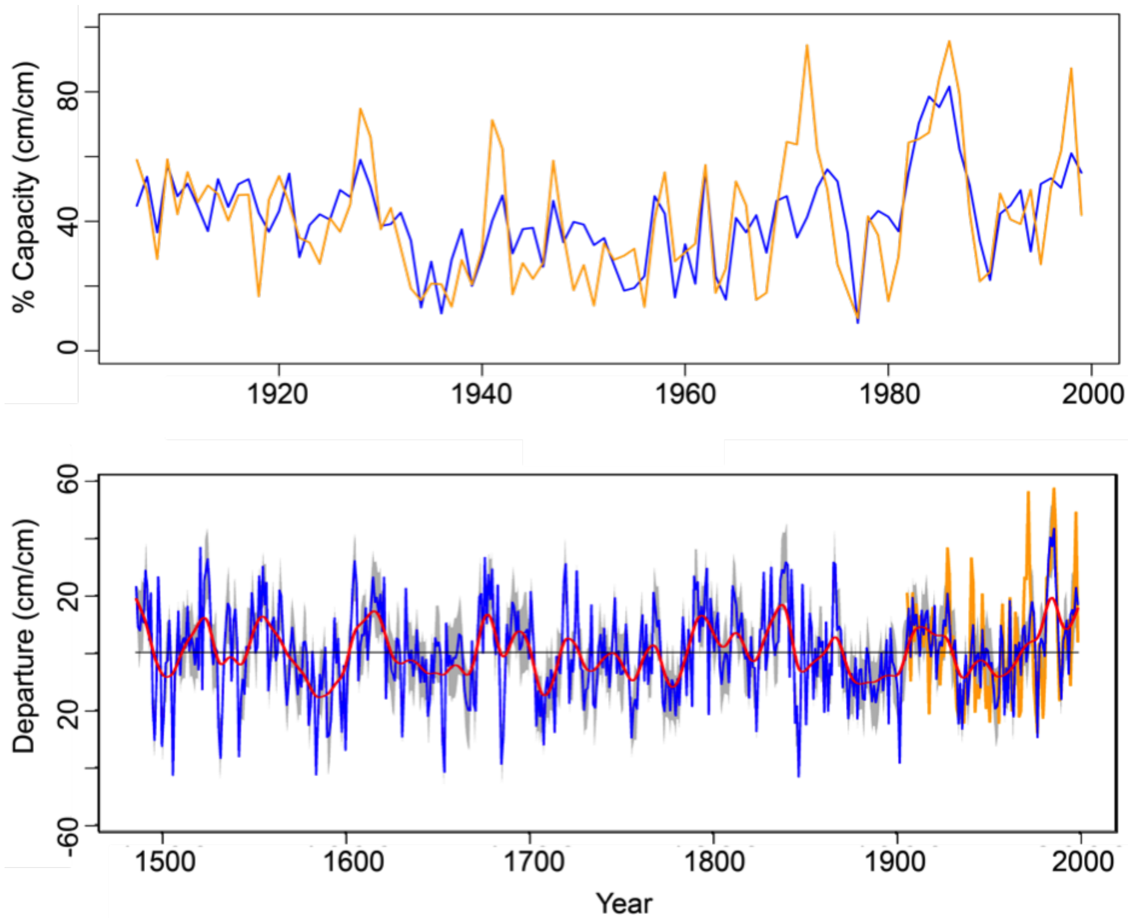

Figure 6 a) McCabe and Wolock (2011) soil moisture (orange) and reconstructed (blue) soil moisture during the reconstruction model calibration period (1906-1998). b) November soil moisture reconstruction for the UCRB (1486-1998). The blue line is reconstructed values, the black line is the long-term mean, and grey shading delineates the upper and lower 95% confidence intervals. Red line is 20-year spline. Orange line is the soil moisture record used during the calibration period.

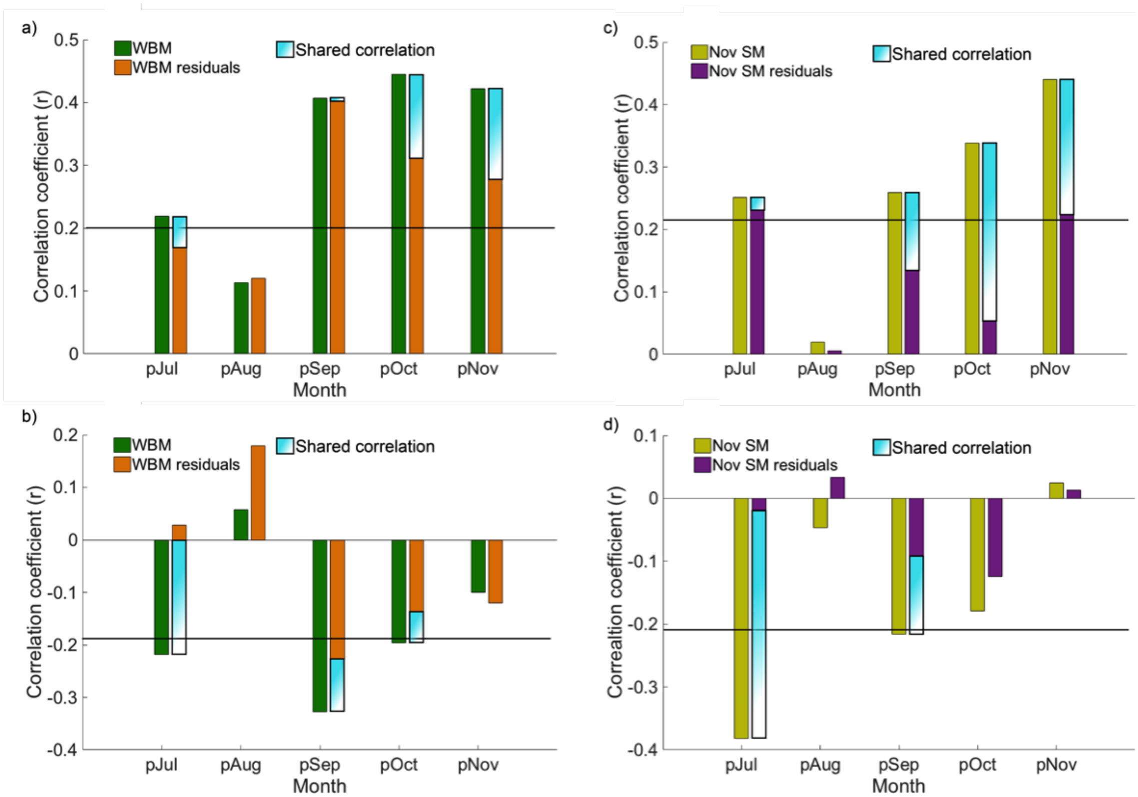

Figure 7 Comparison of correlations before and after removing shared variance with regression: a-b) observed and c-d) reconstructed soil moisture and residuals correlations with antecedent monthly precipitation and temperature. Black line indicates statistical significance ($p < 0.05$) for precipitation (a and c) and temperature (b and d). Gradient bars indicate the assumed shared climate information removed during regression.

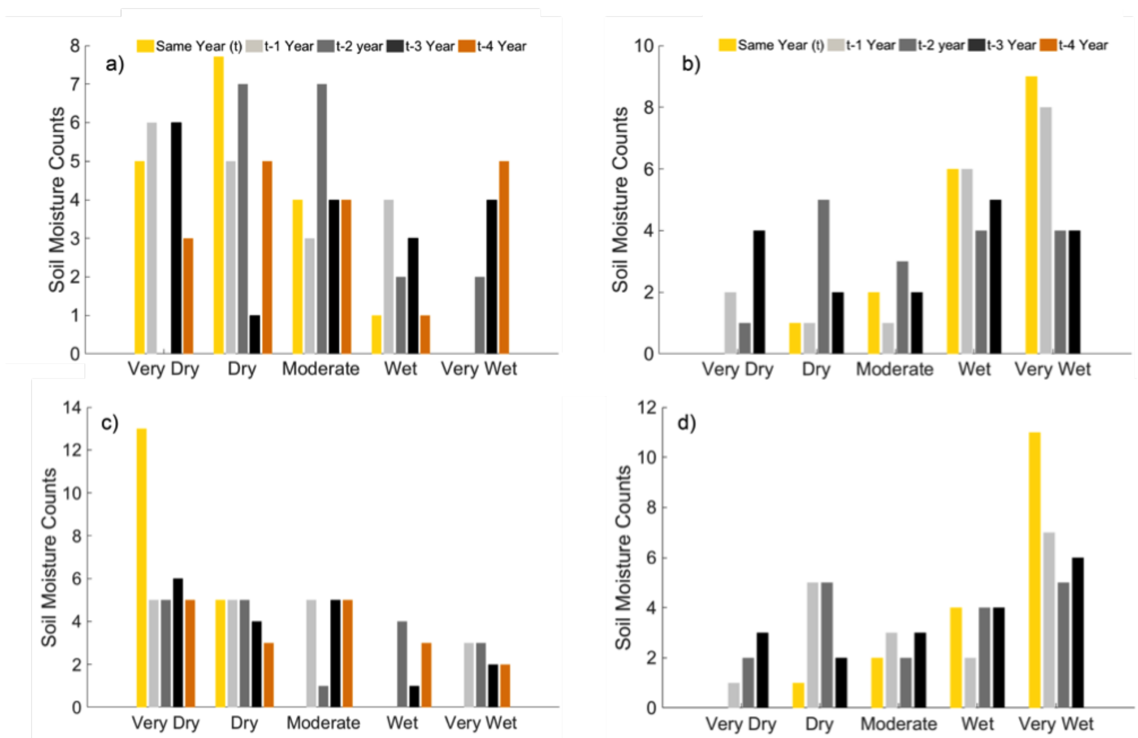

Figure 8 Counts of WBM November soil moisture in t-x years* from 20th percentile (low flow) or 80th percentile (high flow) gaged streamflow years (1906-1998). a) low-flow years, and b) high-flow years. Counts of reconstructed November soil moisture in t-x years from low (20th percentile) or high (80th percentile) reconstructed streamflow years during the common period (1906-1998). c) low-flow years, and d) high-flow years.

* Year refers to water year. November refers to antecedent November (the November in the water year Oct-Sept).

APPENDIX C

COMPARING TREE-RING BASED RECONSTRUCTIONS OF SNOWPACK VARIABILITY AT DIFFERENT SCALES

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Abstract

In light of the combination of ongoing drought and limited climate information on the Navajo Nation, Navajo water managers face decision-making challenges complicated by past and future climate uncertainty. This study documents two snowpack reconstruction options developed for Navajo water managers and examines the relevance and usefulness of each of these climate products for the Navajo. Runoff from snowpack in the western United States is a dominant source for water supply. In recent years water availability from snowpack has declined, largely attributed to increasing temperatures in the region. This is a critical issue for many Native American communities who disproportionately rely on local snow-fed water supplies. Working in partnership with the Navajo Nation Water Management Branch, this study addresses Navajo concerns about the amount and variability of snowpack in the Chuska Mountains using two potential snowpack datasets. We used these datasets with tree rings collected in northern Arizona to develop and evaluate reconstructions of Chuska snowpack and their prospective relevance and usefulness to Navajo Water Management Branch decision-making. We found that both reconstructions developed skillfully estimated snowpack, though there were differences and these differences may have meaningful implications for Navajo water managers. The reconstruction that is most representative of Chuska snowpack has less explanatory power than the regionally representative reconstruction, but the Chuska reconstruction effectively captures snowpack extremes and snow drought timing unique to the Chuska Mountains and important to Navajo water management.

Key Words: climate information, snowpack, reconstruction, Navajo Nation, tree rings

1. Practical Implications

Snowpack in the Chuska Mountains is a valuable source of water on the Navajo Nation that is threatened by drought and climate change. Navajo water managers work intensively in the Chuska Mountains to monitor and maintain this important water source. Snowpack monitoring in the Chuska Mountains began in 1985 and, as a consequence, snow records are relatively short. The short-length records make contextualizing current climatic relationships between snow and water in this drought-prone place extremely difficult. Without consistent, long-term climate information related to snowpack, Navajo water managers face significant challenges with anticipating impacts from climate variability. Informal conversations with Navajo water managers about recent snowpack declines and earlier seasonal runoff provide accounts of the economic and cultural consequences of these declines. At the same time, Navajo water managers are searching for quantitative documentation of historical changes in snowpack that supplements, informs, corroborates, and supports existing tribal knowledge, and can in turn help to guide decision-making among local resource managers.

The Navajo Nation Water Management Branch initiated this collaborative research. It was an ongoing, interactive approach to climate information production that was guided by the Navajo Nation, that was driven by their management needs, and intended to inform Navajo planning for water sustainability in times of drought and in the face of projected warming. The relationships developed through this process were critical to ensure that snowpack reconstructions were relevant and useful. These relationships also support future endeavors to forge integrated science-water management partnerships with tribal governments.

Tree-ring based climate reconstructions require two main components, 1) a calibration dataset of the climate variable that satisfies a minimum-length convention to produce a robust reconstruction model and minimize model errors (minimum of 30 years), and 2) tree-ring chronologies that exhibit a statistical relationship to the variability in the climate data. Intuitively, Chuska Mountain snow records (1985-2015) should be the dataset used to calibrate a Chuska Mountain snowpack reconstruction model. But, short Chuska Mountain snow records raised questions about the scientific robustness of a reconstruction generated using such limited data. For this reason, we developed two tree-ring based snow water equivalent (SWE) reconstructions. One reconstruction is calibrated on the SWE record from the Chuska Mountains. The second reconstruction is calibrated on a longer SWE record, Williams Ski Run (1967-2015), from the San Francisco Peaks approximately 250 km to the southwest. The Williams Ski Run SWE data is representative of SWE conditions in the region, which includes the Chuska Mountains. We then compare the resulting reconstructions in terms of reconstruction skill and model validation, and the ability of the snowpack estimates to replicate observed snowpack data characteristics. The reconstruction calibrated on the Chuska snow record better matched the details of snowpack variability in the instrumental record, but generally failed to capture the magnitude of extremes. The Williams Ski Run reconstruction captured a broader range of regional snowpack variability, but it missed local low-snowpack intervals specific to the Chuska Mountains. Knowing these trade-offs allows Navajo water managers to determine what climate information contained within the reconstruction is most useful for their immediate decision-making.

The research reflects on the usefulness of climate information given that use-inspired science is complex, time-intensive, and must enable knowledge production that is beneficial to, and reflects the concerns and needs of, the information user. What makes this kind of research difficult is that the value attributed to research relevance can be different for the researcher versus the user of the information. Further, the research must be believable, trusted, and readily usable by Navajo water managers in order to adequately meet their needs. We found that this dual-pronged approach of 1) Navajo directed research objectives, and 2) comparisons of climate services products according to scale begins to address the gap in climate information on the Navajo Nation while also producing information that specifically addresses locally relevant questions.

2. Introduction

Gaps in actionable climate information for water management decision making often arise from spatially sparse hydroclimatic data and short-duration climatic records. These factors have historically limited assessments of climate vulnerability by resource managers on the Navajo Nation in the southwestern U.S., a sovereign Native American reservation with over 300,000 residents (Novak, 2007; Ferguson et al., 2011; Redsteer et al., 2013; Tsinnajinnie et al., 2018; Tulley-Cordova et al., 2018;). There is a need for customized climate information on the Navajo Nation that integrates Western science and indigenous knowledge in ways that are beneficial to both knowledge systems (Redsteer et al., 2010; Chief et al., 2016), that can be readily used in climate-related decision-making (Yazzie et al. 2019), and that ensure that relevant and trusted climate information is the outcome of a process that considers the concerns and perspectives of the user of the information (Cash et al., 2002; Cash and Buizer, 2005; McNie, 2013).

Climate services is emerging as a functional framework that capitalizes on diverse expertise (Brasseur and Gollardo, 2016), on recent scientific advances, and on the co-production of knowledge (Bremer et al., 2019) to produce user-relevant climate products to support decision-making at various scales (e.g. Cortakar et al., 2016). Definitions of climate services incorporate key components of climate knowledge production including the timely availability and customization of climate information, efficient transfer and translation of that information, and guidance or counseling on using the information to support climate change adaptation, mitigation, and risk management (Brasseur and Gallardo, 2016). According to Brasseur and Gallardo (2016) lacking or insufficient climate services components, including the lack of user-relevant products offered by the scientific community, present challenges to the success of climate services. In this study, we focus on improving one component of climate services, the development of relevant and usable climate information at the local scale (heretofore referred to as climate information). Research demonstrates that collaborative development of climate information is more likely to result in useful science (Jasanoff and Wynne 1998; Jasanoff, 2004; Lemos and Morehouse 2005; van Kerkhoff and Lebel 2015). Useful climate information is therefore most likely achieved when decision makers define the problem and the desired climate product (Clark, 2002), and when the users of the information participate in its production (Lemos and Morehouse, 2005; Tall and Njinga, 2013; Lemos et al., 2014; Wall et al., 2017).

The Navajo Nation has been in a state of drought emergency since 2011 (Cozzetto and Nania, 2014) and severe drought has affected the area since about 1999 (Redsteer et al., 2011; Crimmins et al., 2013). The drought is negatively affecting crops, food supplies, water storage, economic conditions and ecosystem services (Ferguson et al., 2016; El-Vilaly et al., 2018), and is often raised by tribal members and especially elders as an unusually long-lasting problem (Ferguson et al., 2011; Redsteer et al. 2011). Local observational climate data to address concerns over the recent drought are either lacking or too short to estimate long-term changes in trend or variability. For example, assessments of snowpack and snow water equivalent (SWE) measurements in the Chuska Mountains of the Navajo Nation have been ongoing since the 1980s, but this time period has been insufficient to reveal a discernable trend due to drought or climate change (Tsinnajinnie et al., 2018). The lack of trend in SWE is inconsistent with observations of declining surface waters and streamflow across the Navajo Nation (Redsteer et al., 2011) and with reports of drying snow-fed lakes in the Chuska Mountains. These reductions in surface waters are likely connected to SWE and its variability, however it is difficult to distinguish the role of snowpack decline versus increasing temperature with short instrumental records, prompting the need for long-term data for snowpack levels in the Chuska Mountains.

Driven by the need to plan for and adapt to climate change, water managers of the Navajo Nation invited us to help in understanding local variability in climate and water resources. They identified the amount and variability of snowpack in the Chuska Mountains as a key concern. The Navajo Nation encompasses over 70,000 km^2 in the Four Corners region of the American Southwest (Figure 1). Much of the reservation is high-desert grassland, typical of the Colorado Plateau, but on the eastern part of the reservation the Chuska Mountains reach nearly 3,000 m elevation with a winter precipitation regime that is dominated by snow. The Chuska Mountains provide most local surface water to the eastern portion of the Navajo Nation (Harshbarger and Repenning, 1954; Wright 1964; Garfin et al, 2007) and are the headwaters for several perennial creeks that feed multiple reservoirs and river systems, such as the Little Colorado River, the San Juan River, and in Canyon de Chelly. Navajo community members on the eastern reservation are strongly reliant on snow-fed surface water for livestock, fishing, and agriculture (Crimmins et al., 2013; Wright, 1964), diverting water resources to feed small-scale irrigation structures to support traditional farming communities (Harshbarger and Repenning, 1954). These systems may be threatened by increased temperature and projected shifts in cool-season precipitation toward sporadic snow accumulation and earlier spring melt (Mote et al., 2006; Li et al., 2017).

To better understand fluctuations in snowpack over time, we utilize tree rings to reconstruct snowpack for the Navajo Nation. The hydrological and biological basis for using tree rings is that the growth of southwestern U.S. montane conifers living on well-drained, south-facing slopes is controlled by winter precipitation (Fritts, 1976; Grissino-Mayer, 1997; Woodhouse, 2003; Touchan et al., 2010; Pederson et al., 2011; Faulstich et al., 2013). Water, which arrives in winter in these locations, controls tree growth in the subsequent growing season through snowmelt entering the root zone in late spring and early summer (St. George and Ault, 2014).

The goal of this study was to work in partnership with the Navajo Water Management Branch (NWMB) to place NWMB snowpack data into a centuries-long context using tree rings, producing relevant and useful climate information for NWMB managers. In our preliminary work, we found that a SWE reconstruction based solely on Chuska Mountain data fell short of our expectations for generating a robust statistical model (e.g., the length of the calibration dataset is only 30 years in length). We also aimed to capture as much natural variability as possible from the calibration dataset. In the southwestern United States, where precipitation variability is high, short calibration datasets are likely to miss important extremes. Therefore, we generated a second reconstruction from regionally available data that better met standards for skillful reconstructions, but that may have traded its local usefulness in the process. Here, we evaluate the relevance and usefulness of these two reconstructions in terms of 1) the statistical robustness (reconstruction skill and validation), and 2) the replication of observed snowpack characteristics most meaningful to the NWMB. We further use the multi-century reconstructions to assess the magnitude and duration of low snowpack periods in the 20th and 21st centuries in a longer-term context.

3. Materials and Methods

3.1. Data

3.1.1. Snowpack datasets

The NWMB provided us with SWE data for the Chuska Mountains. The data were manually collected from snow course sites near the first and middle of each month from January 1st through April 1st using a snow coring tube and following standard Natural Resources Conservation Service (NRCS) procedures. Snow density was calculated from the mass and volume of snow in the tube. To obtain SWE estimates from the snowpack, snow density calculations were multiplied by snowpack depth. From these data, eight March 1 SWE site records the annual maximum SWE measured on March 1 - were derived spanning 1985-2015 (Figure 2). The eight Chuska Mountains snow sites were then averaged, which we henceforth refer to as the Chuska series. Initial assessment of the eight Chuska March 1 SWE records revealed that coherence between them in the early years (1985-1990) is low, making it difficult to justify using the entire record as a basis for statistical calibration of the reconstruction model. To shorten the record would result in a calibration dataset that does not meet the 30-year threshold for statistical analysis used as a general rule of thumb. Thirty years is a general target and not a requirement for analysis. However, longer calibration datasets allow the reconstruction model to capture a larger range of natural variability, especially longer-term variability (decadal or longer) common in climate signals. To accommodate this shortcoming, we capitalized on an observation that snowpack in the San Francisco Peaks and Mogollon Rim of northern Arizona is closely linked to the Chuska Mountains (~250 km distant) via winter storm tracks (Tsinnajinnie 2011). Therefore, we obtained 26 SNOTEL and snow course site data from northern Arizona as a second set for a potential March 1 SWE calibration. Of these 26 sites, we excluded those with $<$ 30 years of continuous recording, leaving a suite of 17 snow course sites to compare with the Chuska record. Correlations of March 1 SWE from the 17 northern Arizona SWE sites with the Chuska series ranged from $r = +0.67$ (Chandler) to $r = +0.86$ (Happy Jack). Williams Ski Run snow course (WSR; 1967-2015) was longer than our discretionary 30-year threshold, the record contained few zeros or missing values, and had a strong average correlation with the Chuska series ($r = +0.83$). Averaged Chuska Mountains March 1 SWE (CHU; $n = 30$) and Williams Ski Run March 1 SWE (WSR; $n = 48$) were then used for subsequent reconstructions of Chuska Mountain and regionally representative snowpack, respectively.

3.1.2. Tree-ring datasets

The forests of the Chuska Mountains are dominated by ponderosa pine (*Pinus ponderosa*) above 2,300 m, with dry mixed-conifer forests including a large proportion of Douglas-fir (*Pseudotsuga menziesii*) occurring at higher elevations and in cold-air drainages. Lower-elevations are predominantly pion-juniper (*Pinus edulis-Juniperus* spp.) communities. There is a dense network of tree-ring sites in the Four Corners area that includes pion, ponderosa pine, and Douglas-fir chronologies (St. George and Ault, 2014), many of which are available on the ITRDB[1]. We screened the available sites and selected ten tree-ring chronologies from the Navajo Nation and Mogollon Rim/San Francisco Peaks area that had significant correlations ($p < 0.01$) with Chuska SWE and maximum overlap with the calibration. Most of the sites had previous collections (e.g., Dean and Funkhouser 2002; Sheppard et al., 2005) that we updated with collections between 2015 and 2017. Several additional new site collections were made for Guiterman (2016) and we use them in this study. The sites range in elevation from 1,828 m to 2,621 m (Table 1).

Tree-ring samples were collected, mounted, sanded, visually crossdated, and measured for total ring width using standard methods of dendrochronology (Speer 2010). We checked for accuracy in crossdating and performed quality control of the measurement data iteratively in COFECHA (Holmes 1983). Where available, we combined the ring-width series from the original collections with our updated ring-width series. We standardized the ring widths in R using the *dplR* library (Bunn 2008; R core team 2019), employing either a modified negative exponential curve or cubic smoothing spline with a frequency response of 50% at a wavelength of two-thirds the length of the series. Standard chronology statistics were calculated in *dplR*. Each site includes 6-32 trees with chronology statistics that show strong relationships between trees at the site level. The average correlation between trees ($\overline{r}$eff , effective

[1]ITRDB data available at the International Tree-Ring Data Bank, ITRDB; https://www.ncdc.noaa.gov/paleo/treering.html

chronology signal, Cook and Kairiukstis 1990), is above the arbitrary but previously employed lower threshold of 0.5 (Brice et al., 2013) for all but one site that was very close (0.482, San Francisco Peaks Update). All sites show an expressed population signal (EPS) over 0.85, the lower threshold commonly accepted for adequate sample depth for climate reconstruction (Wigley et al. 1984; Briffa and Jones 1990), between the years 1335 and 1666 depending upon the chronology.

3.2. Reconstruction development and analysis methods

3.2.1. Snowpack reconstruction and skill metrics

In preparation for the regression analysis used to reconstruct March 1 SWE, the statistical relationship between the tree rings and the snowpack was evaluated. We used correlation coefficients (r) to test the strength and significance of the relationship between each chronology in the tree-ring network and the snowpack data. Ten chronologies were significantly correlated ($p < 0.05$) with the SWE time series and these were retained in a pool of potential reconstruction model predictors. Normal distribution of the regression variables was verified and met and no significant trend or autocorrelation exists in the March 1 SWE data.

Stepwise multiple linear (least-squares) regression was used to calibrate each reconstruction model on March 1 SWE data (Criteria: Probability-of-F-to-enter $< =$ 0.05, Probability-of-F-to-remove $> =$ 0.10). The R^2 statistic provides a measure of the explanatory power of the model and the F-ratio estimates the statistical significance of the regression equation. The Durbin-Watson (D-W) statistic assesses serial correlation in the model residuals. A D-W statistic at or near 2 indicates zero first order autocorrleation in the regression residuals. The Standard Error of the Estimate (SEE) statistic indicates uncertainty of the predicted values during the calibration period. Leave-one-out cross validation was used to check the reconstruction model performance compared to March 1 SWE observations (Michaelsen, 1987). The leave-one-out process withholds one data point from the calibration period and a prediction is made for that point. This process proceeds iteratively for each value in the calibration period. The validation statistics Reduction of Error (RE) and Root Mean Square Error of the validation (RMSEv) assess the accuracy of the model predictions (Fritts, 1976; Cook et al., 1999). The RE statistic compares the mean square error of the reconstruction to the mean square error of the calibration data average. The RE result indicates if the reconstruction provides more information from the estimates over the validation period than the calibration data mean would provide, and in a skillful model the RE will be nearly equivalent to the R2. The range of the RE statistic is zero to +1, with a positive value indicating skill in the model (a value of +1 meaning perfect skill) (Fritts, 1976).

3.2.2. Runs analysis

We conducted runs analysis to identify multi-year periods of low and of high snowpack in the Chuska Mountains. Following Faulstich et al. (2013), we classified runs periods based upon a threshold of at least two consecutive years above/below the reconstruction mean. Because a single year of above average snowpack may not provide sufficient moisture for the region to recover from several years of below average snowpack (Crimmins et al., 2013), we allowed the consecutive years to be interrupted by no more than one consecutive March 1 SWE year of opposite sign. When defining high snowpack runs, it makes sense to use the same criteria because increased cool-season precipitation is likely to improve wet soil conditions and recharge of local water resources despite a single year below average (Redsteer et al., 2010; Crimmins et al., 2017). Assessing drought in the Southwest is sensitive to decisions of runs thresholds (Meko et al., 1995). For example, without our exception rule allowing one season of opposite sign in a run, the 1950s drought (1950-1964 in the Williams Ski Run results) would be two separate dry periods (1950-1951, 1953-1964), thus minimizing the duration and magnitude of the drought in our interpretation of results and not accurately representing the impacts of the dry period. Because of the high interannual variability in this region, and because of the decadal scale persistence above or below the mean in the smoothed series, we have chosen to identify runs periods in this way. Duration (number of consecutive years), magnitude (cumulative deficit), and intensity (magnitude divided by duration) for each drought or pluvial event were ranked using the method described in Faulstich et al. (2013). After assigning a rank for each measure, the ranks were summed for a total score. The total scores per event were then ranked to establish the most extreme drought and pluvial periods.

3.2.3. Decadal-scale variability, ranked average deficits, and extreme years

To better understand decadal-scale variability in the reconstruction, a 20-year cubic smoothing spline was calculated and overlaid on the annual March 1 SWE reconstructed series. The spline was used to identify periods when the smoothed series remain above or below the long-term mean and to assess the distribution through time of prolonged above or below average snowpack. The unsmoothed reconstruction and the smoothed reconstruction were converted to a departure series by subtracting the long-term reconstruction mean. The departure series were ranked. The ten driest individual departure years by rank were evaluated in terms of largest deficit in a single year in the reconstruction. The five lowest non-overlapping periods in the smoothed series were centered on lowest value in the smoothed series and averaged over the period for which values were negative.

4. Results

4.1. Evaluation of the calibration datasets

The calibration datasets analysis revealed important differences between Chuska (CHU) and Williams Ski Run (WSR) observations. The CHU March 1 SWE mean is 18.69 cm, with a standard deviation and variance of 8.32 and 69.19, respectively (Table 3). The WSR March 1 SWE mean is 21.33 cm, with standard deviation and variance of 12.86 and 165.53, respectively. Series ranks for CHU and WSR show that 2006 was the driest year in the instrumental record for each of the two SWE sites. The ranks of the remaining four years are not the same between the two series. The years 2015 and 1996 are both ranked top five, however 2015 was third (CHU) and second (WSR), and 1996 was fourth (CHU) and fifth (WSR). The extremely dry year across the southwestern U.S., 2002, ranked third driest in the WSR record but it did not rank top five in the CHU record. After standardization, the overall intensity of the 2000s drought (1997-2007) in CHU was -0.250 cm and the overall intensity of the same drought in WSR was -0.527 cm. The WSR running total (magnitude) for the 2000s drought was 47% drier than the CHU magnitude. The percent of average over these years was 88% of average in CHU and 64% of average in WSR.

4.2. Chuska local snowpack reconstruction

Stepwise regression identified one tree-ring chronology collected in the Chuska Mountains from *P. ponderosa* at Small Twin Canyon (STC) to predict Chuska Mountains March 1 SWE (CHU). The final reconstruction model is:

$$CHU = 4.035 + 0.534(STC) \tag{C.1}$$

The CHU model explains 41% of the variance in Chuska March 1 SWE in the 30-year calibration period (1985-2014, Table 2). The F-ratio indicates that the regression equation is statistically significant. The RE (0.41) and RMSEv (6.198) values are comparable to their respective calibration statistics, $R^2 = 0.41$ and SEE = 6.523, showing that this model has skill in estimating Chuska SWE during cross-validation (Table 2, Figure 3). The sign test demonstrates that the direction of observed and estimated departures from the instrumental mean agree more often than would be expected by chance alone. Analysis of reconstruction residuals revealed no violation of regression assumptions. Reconstruction residuals are normally distributed, show no significant trend or changes in variance with time, and no significant autocorrelation. Reconstructions tend to underestimate extreme years, and this is demonstrated in the years where observation values are higher or lower than reconstructed values. However, this reconstruction replicates some extreme values found in the calibration series (i.e. 1988, 1990, 1995, 1996, 2002, 2006). The model also underestimates total variance during the calibration period. The full reconstruction spans from 1656 to 2014.

4.3. Williams Ski Run snowpack reconstruction

Stepwise regression identified two tree-ring chronologies collected in the Chuska Mountains from *P. ponderosa* and *P. menziesii* at Small Twin Canyon (STC) and Spider Rock (SSR) as the best predictors of Williams Ski Run March 1 SWE (WSR). The final reconstruction model is:

$$WSR = -3.113 + 0.363(STC) + 0.381(SSR) \tag{C.2}$$

The model explains 47% of the variance in Williams Ski Run SWE in the 41-year calibration period (1967-2014, Table 2). The F-ratio indicates that the regression equation is statistically significant. The RE (0.40) and RMSEv (9.870) values are comparable to their respective calibration statistic, $R^2 = 0.47$ and SEE = 9.548, showing that this model skillfully estimates WSR SWE during cross-validation (Table 2, Figure 3). As with the CHU model, tests show no violation of regression assumptions. Sign test results demonstrate significant agreement between the WSR calibration series and the reconstruction during the calibration period. The WSR model best captures below-average observed values in the second half of the calibration period, rather than above average observed values in the same interval. The length of the reconstruction is limited by the shortest chronology that contributes to it. Cutoff years for robust chronologies to be used in the reconstruction are 1694 (STC) and 1654 (SSR). The STC chronology (1656-2014) reached an EPS value of 0.85 at 1694, and thus the full reconstruction spans from 1694 to 2014.

4.4. Analysis of the reconstructions

The smoothed series reveals that the duration of multi-year periods of low and high SWE varies similarly in both reconstructions across the three centuries (Figure 4). In the early-to-mid 1700s, periods of low snowpack occurred between long intervals of above-average snow. The first half of the 19th century is dominated by a long-duration below-average snowpack covering the 1820s. In the second half of the 19th century, low snowpack periods were of shorter duration frequently interrupted by similar length or longer wet intervals. The 20th century has less frequent and long dry or wet periods.

Runs analysis highlights the duration of persistent snowpack conditions and reveals periods of extremely high and low snowpack (Table 4). The duration of high snowpack periods from the CHU reconstruction in the top five high snowpack intervals ranges between 11 and 18 years. The highest ranked run of high snowpack occurred in 1915-1932 (18 years). This period was also the longest run of high snowpack in the WSR reconstruction (39 years). Of the four highest ranked high CHU snowpack years, three periods occur in the 1900s. Low CHU snowpack in the top 5 runs periods ranges between 8-28 years, whereas the top 5 low snowpack periods in the WSR reconstruction range from 8 to 17 years in duration. The extended low CHU snowpack period (1950-1977, 28 years) is the longest, and ranked most severe in all the runs. The most recent drought (1999-2006) ranks number six among the ten lowest snowpack periods in the CHU reconstruction. In the WSR reconstruction, the duration of the top five high snowpack periods ranges between 5 and 39 years. The highest ranked period of high snowpack occurred in 1718-1727 (10 years). The longest run of high WSR snowpack occurred in 1907-1945 (39 years). Of the four highest ranked high snowpack years, three periods occur in the 1700s. The extended low WSR snowpack period (1818-1834, 17 years) is the longest, and ranked most severe. The low snowpack period (1950-1964, 15 years) is also present in the WSR reconstruction runs, ranked number 4 in severity in the ten ranked periods. The most recent drought (2000-2007) ranks number five among the ten lowest snowpack periods in the WSR reconstruction.

Extremely dry single years occur frequently during longer deficit periods (Figure 5). The single driest year in the CHU reconstruction is 1729 and falls within a longer deficit ranking in the top five deficit periods in the reconstruction. The single driest year in the WSR reconstruction is 2002. Other top ten dry years such as 1822, 1900-1904, 1951, and 2006 (CHU only) coincide with extended, severe average deficits (based on the 20-year spline) in the early 1800s, the early 1900s, and the mid-20th century. Other dry years in the WSR reconstruction, such as 1700, 1847, 1861, 2002, and 2006 are not part of a longer deficit in consecutive years relative to other periods in the long-term record. The year 1847 is the single individual year in the CHU reconstruction that is not within a longer deficit period.

5. Discussion

This study aimed to produce the most relevant and useful information about multi-century snowpack variability in the Chuska Mountains for the Water Management Branch of the Navajo Nation. In so doing, we assess the trade-offs between the use of a relatively short, but locally-relevant snowpack instrumental record versus a longer, but potentially less locally-relevant instrumental record in developing a tree-ring based reconstruction of snowpack variability. In a comparison of the two observed snowpack records and reconstructions, we i) assess the statistical characteristics and extremes of the instrumental data, ii) evaluate the ability of each reconstruction to reflect means and extremes in the instrumental data, and iii) compare the ability of the reconstructions to capture known drought episodes. To assess the reconstructions as usable and relevant climate information, we i) evaluate the SWE reconstruction implications for local water resources and ii) evaluate the collaborative development of climate information.

5.1. Evaluating the instrumental records for March SWE

Our findings show similarities and differences in the observed snowpack volume and variability between the Chuska Mountains (CHU) and the Williams Ski Run (WSR). Though the 30-year averages between datasets are similar, and the datasets are highly correlated, the strong correlation arises from synchrony in sign, not the magnitude of above-average or below-average snowpack, which is generally larger at WSR, especially in extreme years. The only exception is 2006, for which both datasets showed near-zero March 1 SWE. The 2006 winter had the lowest snowpack in the 1985-2015 record at both sites.

The rather extreme 2000s drought (1997-2007) further highlights the differences between the instrumental datasets. Overall, using just the WSR record, one would conclude that the drought was far more severe than if using the CHU record. This is exemplified by the magnitude of 2002 low snowpack at Williams Ski Run, which is half of the SWE observed in the Chuska Mountains for that year. These local differences underscore the importance of evaluating observed datasets used for use-inspired research when stakeholders need science specific to locally driven climate questions.

5.2. Evaluating the tree-ring reconstructions

Our tree-ring reconstructions skillfully estimate observed March 1 SWE, but differences in their magnitudes and signs may prove meaningful to NWMB. The Williams Ski Run (WSR) reconstruction shows greater explanatory power ($R^2 = 0.47$), so it provides more accurate information about snowpack variability from one year to the next. The validation statistics for each reconstruction indicate that each model demonstrates skill in estimating values not included in the instrumental data during the leave-one-out validation process (Table 2). The CHU reconstruction is relatively effective at capturing individual extremely dry years found in the calibration data. It does not capture the severity of 2006 as well as the Williams Ski Run reconstruction does, even though 2006 is extremely dry in both instrumental data series. But, the CHU reconstruction does capture the dry year 1990 in the Chuska Mountains, which was only an average year in the WSR instrumental data. In southwestern US states such as Arizona, Utah and Colorado, 2002 is associated with much below normal to record-low precipitation (e.g. Breshears et al., 2005; Williams et al., 2015). Reflecting this, 2002 was a snow drought year in both the WSR calibration and reconstruction series. The Chuska Mountains also experienced snow drought in 2002, but as is consistent with other local instrumental records observed in New Mexico, 2002 was not exceptionally dry in the Chuska Mountains. The reconstruction also reflects this difference.

Decadal to multi-decadal periods of low snowpack found in the reconstructions are consistent with other studies in the region. Three severe SWE droughts in the WSR reconstruction 1728-1744, 1818-1834, and 1893-1908 also rank highly in a cool-season precipitation reconstruction for the Four Corners (Faulstich et al. 2013). The second highest-ranking dry WSR snowpack period (1728-1744) and the highest-ranking CHU snowpack period (1729-1742) are also the highest-ranking cool-season drought in the Faulstich study. This drought corresponds with social upheaval in northwestern New Mexico and with a long-duration dual-season drought (cool-season and warm-season dry intervals occurring in the same year) in the region (Faulstich et al., 2013). Other nearby research supports our results showing that droughts prior to the instrumental record have been more intense or longer-lasting than dry periods of the 20th century (Woodhouse and Overpeck, 1998; Novak, 2007). Reconstructed October-July precipitation in the southern Colorado Plateau reveals coherent cool-season deficits between the Williams Ski Run record and a broader record for north central Arizona and south central Utah (Salzer and Kipfmueller, 2005). The cool-season drought beginning in 1818 is the most severe cool-season drought in the southern Colorado Plateau, ranked the most severe snow drought in the WSR reconstruction, and it is extremely dry in the CHU reconstruction. Two other low snowpack periods ranking among the ten lowest in both reconstructions also rank among the most severe cool-season droughts in the southern Colorado Plateau region, 1890s-1900s and 1750s-1760s. While rank and duration of dry periods compared between these studies are generally consistent, differences may be attributed to i) differing drought-period thresholds defined in each study, ii) that this study focuses only on SWE rather than other climate variables, or iii) the existence of large variation in local SWE signals. Some variation in rank and duration between studies is expected, but our SWE reconstructions align with previous work demonstrating their skill to represent past SWE.

5.3. SWE reconstruction implications for local water resources

The CHU and WSR reconstructions in this study show that past snow droughts were of greater magnitude than severe snow droughts of recent memory. Further, extremely dry individual years punctuate multi-year drought periods in a way that has not been recognized from instrumental data alone and that can have a large influence on the overall intensity of a given drought. Extremely dry periods present in the paleo record, often more severe than what has been experienced in the instrumental record, coincided with impacts to human civilization (Cook et al., 2007).

These impacts include societal disruptions in the Navajo region during periods of coinciding cool- and warm-season drought (Faulstich et al., 2013). Both reconstructions reveal the presence of extremely dry years embedded in longer dry periods, some of which coincide with tangible and documented impacts to Navajo water resources in the 20th and 21st centuries.

Despite its significant recent impacts, the 2000s drought is ranked as only the sixth driest run in the CHU reconstruction, is only half or less the duration of the 1700s, early 1900s, and 1950s droughts, and has a magnitude (running SWE total of years with snowpack below the mean) 50-75% lower than these others. In addition to deficits in the 2000s, drought impacts experienced during the 1950s still resonate with Navajo living at the time (Novak, 2007; Redsteer, 2011). Although the year 1951 is among the driest individual years in both reconstructions, and the lengthy mid-century dry interval is of notably long duration, the 1950s are rivaled and exceeded by other dry snowpack periods when compared to the past 300 years. The magnitude of the 1950s drought in the WSR reconstruction is only 67% of the magnitude of the highest-ranking droughts in the same record, 1818-1834 and 1728-1744. By contrast, the CHU reconstruction captures a longer and larger magnitude 1950s drought in the Chuska Mountains, rivaled only by the early 1900s drought.

Now with a record of Chuska snowpack variability over the last three centuries, the relationship between water scarcity and snow drought should be considered. Extreme water scarcity experienced since the early 2000s on the Navajo Nation (Redsteer, 2011) is largely attributed to warming and drying, but water scarcity may have been damped by concurrent years with moderate to high snowpack. Crimmins et al., (2017) showed a shift during the 2000s in 50% cumulative total annual precipitation to later in the spring relative to the 1950s. This suggests that reduced snowpack and more precipitation falling as rain may not moderate drought impacts in the Chuska Mountains. Persistent snowpack supports recharge of surface water, groundwater, and springs (Ferguson et al., 2011; Lani Tsinnajinni, personal communication), but years or decades when snowpack remains low and more likely to disappear earlier in the spring may worsen already persistent water shortages, such as during the 1820s and 1900s. Impacts of such changes could be severe for the Navajo.

5.4. Assessment of the collaborative development of climate information

This research followed a process meant to lend legitimacy (Cash et al., 2002) to the production of climate information, and to develop a final SWE reconstruction product that is beneficial and readily usable to Navajo water managers. Our efforts to develop climate information for NWMB were individualized to NWMB practices and targeted at the local scale (McNie et al., 2007; Lemos et al., 2012). We worked actively with Navajo water managers to identify research needs. Through collaboration we formulated the research question that effectively aligned with NWMB needs, and were also within the capabilities of the researchers. Over a two-year period, there were eight in-person visits to the NWMB offices, with four expeditions to the Chuska Mountains. These visits consisted of meetings and discussions with NDWR staff about their water resources questions and involved brainstorming approaches to work together to try to answer those questions. During initial meetings we articulated data acquisition requirements and conducted on-site Navajo library research. During subsequent meetings, we visited remote Chuska snowcourse and SNOTEL sites and conducted a nearly complete north-south survey of Chuska Mountain lakes. Meetings were organized with Navajo researchers who were concurrently conducting climate research in the Chuska Mountains, and with Margaret Hiza-Redsteer, United States Geological Survey staff scientist, who investigated increasing aridity in northeastern Arizona (Redsteer et al., 2011). We held interim workshops and meetings to ensure water managers were aware of the research, methods, direction, and progress. These meetings were useful to assess whether the evolution of the research answers the research questions, and served to iteratively exchange results and ramifications of the results. We rigorously vetted possible reconstructions and followed scientific protocols that assure credibility (McNie, 2013). For example, we followed standard dendrochronological procedures and statistical tests, as well as presented the research to the scientific community. Results and data were transferred to the community by way of the NWMB for their specific use (Chief et al., 2018) and through a results webpage[2]. The climate information generated through this research was presented to the tribal community at large in two settings, i) at the 2017 Navajo Nation Department of Natural Resources conference near Flagstaff, Arizona and ii) an organized tour of the Navajo forest, where a cadre of scientists joined natural and cultural resource managers as well as community members from the Navajo Nation to discuss aspects of climate vulnerability, including issues and opportunities in water management for the Chuska Mountains.

It can be difficult to identify and qualify the effectiveness of climate services after the science products are provided to the decision-making partners. Despite high levels of interaction with NWMB, organizational and material limitations on both the Navajo and academic partners could have constrained the scale of the application of our reconstructions (Wall et al., 2017; Lemos et al., 2012). For this reason, we use an output and impact framework from Wall et al. (2017) to evaluate the potential effectiveness of the two-snowpack-reconstruction approach used in this study. Our outputs include two new data series representing local and regional scale reconstructions. This output provides statistically robust and long-term climate information that did not exist prior to the study. While working closely with NWMB agency representatives we were able to increase comprehension and refine relevance of this information in real-time. The researchers received direct confirmation from the representatives concerning their perception of their own understanding of the information. The two-series reconstruction approach should motivate the users of our

[2] Becky Brice, Chuska Mountains, Navajo Nation - https://npsbec.wixsite.com/coplathydroclim/chuska-mountains-navajo-nation

climate information to evaluate results in terms of application and relevance, though this cannot be confirmed at the time of this publication. The comparative approach used here also provides a richer context for the climate information that situates local snowpack variability in the surrounding climate and illuminates limitations and benefits of employing various information resources at different scales. The two-reconstruction approach suggests the need for additional similar research in a region where instrumental climate information is historically limited. This research also highlights the need for strategic research collaborations between tribes and academic institutions informed by climate services research and that increase the range of uses of climate information derived from user-specified research questions (Lemos et al., 2012).

6. Conclusion

This study is an example of use-inspired science driven by the expressed need for climate information in the context of severe drought and declining water resources on the Navajo Nation. We worked in partnership with the Navajo Water Management Branch (NWMB) to place snowpack data into a centuries-long context using tree rings while producing climate information that is both relevant and useful for NWMB managers.

Snowpack information developed from two reconstruction models - one localized to the Chuska Mountains (CHU) and one for the Williams Ski Run (WSR) reveal differences in the ability of the individual reconstruction models to capture the variability present in the observed data. The WSR reconstruction was able to demonstrate greater accuracy in estimating SWE values. These reconstructions also reveal some differences in the timing and intensity of individual extreme years and drought intervals. However, despite concerns about the shorter length of the CHU record (30 years) the reconstruction model calibrated on this data was still able to capture 40% of the variance in SWE in the Chuska Mountains, was generally consistent in reflecting SWE variation in the respective observed record, and more effectively captured the duration, magnitude and timing of recent droughts having an effect on the people living in the region. For these reasons, our results suggest that the local Chuska Mountain reconstruction has greater potential to be relevant and useful climate information to the NWMB.

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Tables

Table 1 Tree-ring site chronologies used in the predictor pool (statistics described in the text).

Site*	Site Name	Speices	First Year	Last Year	Number Cores*	Number Trees*	Elev. (m)	Collection Date	Update Collectors	Publication
SPU	San Fancisco Peaks Update	PSME	1763	2016	46	24	2036	Jan-17	Brice; Sheppard; Arizpe	*Salzer, M. W., & Kipfmueller, K. F. (2005). Reconstructed temperature and precipitation on a millennial timescale from tree-rings in the southern Colorado Plateau, USA. Climatic Change, 70(3), 465-487.*
SMU	Slate Mountain Update	PIPO	1590	2016	60	31	2714	Jan-17	Brice; Sheppard; Arizpe	*Meko, D. M., & Hirschboeck, K. K. (2008). The Current Drought In Context: A Tree-Ring Based Evaluation of Water Supply Variability for the Salt-Verde River Basin Final Report. Salt River Project, Tucson, Ariz.*
RMU	Robinson Mountain Update	PIPO	1621	2016	46	24	2130	Jan-17	Brice; Sheppard; Arizpe	*Meko, D. M., & Hirschboeck, K. K. (2008). The Current Drought In Context: A Tree-Ring Based Evaluation of Water Supply Variability for the Salt-Verde River Basin Final Report. Salt River Project, Tucson, Ariz.*
SCU	Sunset Crater Update	PIPO	1837	2015	25	17	2127	Jan-17	Brice; Sheppard; Arizpe	*Sheppard, P. R., May, E. M., Ort, M. H., Anderson, K. C., & Elson, M. D. (2005). Dendrochronological responses to the 24 October 1992 tornado at Sunset Crater, northern Arizona. Canadian Journal of Forest Research, 35(12), 2911-2919.*
OCW	Oak Creek Wash	PSME	1200	2015	29	19	2325	Jan-17	Guiterman	*Guiterman, C. H. 2016. Climate and human drivers of forest vulnerability in the US Southwest: Perspectives from dendroecology. PhD Thesis. University of Arizona.*
SRD	Spider Rock Douglas Fir	PSME	1636	2014	12	6	1980	Jun-16	Guiterman	*Guiterman, C. H. 2016. Climate and human drivers of forest vulnerability in the US Southwest: Perspectives from dendroecology. PhD Thesis. University of Arizona.*
STC	Small Twin Canyon	PIPO	1656	2014	10	10	2152	Jun-16	Guiterman	*This study*
SRO	Spider Rock Overlook	PIED	1601	2015	62	32	2134	Jun-16	Guiterman	*Guiterman, C. H. 2016. Climate and human drivers of forest vulnerability in the US Southwest: Perspectives from dendroecology. PhD Thesis. University of Arizona.*
SSR	South of Spider Rock	PSME	1396	2014	10	10	1980	Jun-16	Guiterman	*Guiterman, C. H. 2016. Climate and human drivers of forest vulnerability in the US Southwest: Perspectives from dendroecology. PhD Thesis. University of Arizona.*
DCC	Defiance Cross Canyon	PIPO	1340	2015	52	21	2159	Jun-16	Guiterman	*Guiterman, C. H. 2016. Climate and human drivers of forest vulnerability in the US Southwest: Perspectives from dendroecology. PhD Thesis. University of Arizona.*

*includes original plus updated collections

Table 2 Stepwise regression model results for the two Mar 1 SWE reconstructions, a) Chuska Mountains (CHU) and b) Williams Ski Run (WSR).

a)
CHU Reconstruction

Reconstruction	R^2	Adj R2	SEE	Durbin-Watson	F-ratio
	0.41	0.39	6.523	1.742*	19.18**
LOO Validation***		RE	RMSE	Sign Test (hit/miss)	
		0.41	6.198	25/4**, *N*=29	

b)
WSR Reconstruction

Reconstruction	R^2	Adj R2	SEE	Durbin-Watson	F-ratio
	0.47	0.45	9.548	2.051*	20.17**
LOO Validation***		RE	RMSE	Sign Test (hit/miss)	
		0.40	9.870	34/13**, *N*=46	

*H0: Zero first order autocorrelation in residuals. Accept; prob level 0.01
**significant at p<0.05
*** Leave-one-out cross validation description in the text

Table 3 Instrumental (CHU = 1985-2015; WSR = 1967-2014) and reconstruction statistics during each calibration period, and for the full reconstructions (CHU = 1656-2014; WSR = 1694-2014).

	N	Mean	Minimum	Maximum	Range	Std. Deviation	Variance
CHU observed	30	18.69	0.91	38.24	37.33	8.32	69.19
CHU Recon (cal period)	30	18.69	10.26	29.51	19.25	5.31	28.14
CHU Recon	359	19.00	10.26	32.10	21.84	4.68	21.912
WSR observed	48	21.33	0.00	49.78	49.78	12.86	165.53
WSR Recon (cal period)	48	21.33	-2.67	37.87	37.87	8.70	75.68
WSR Recon	359	21.97	-2.67	46.32	46.32	9.32	86.829

Table 4 Runs analysis for low snowpack (left) and high snowpack (right) periods in the CHU (top) and WSR (bottom) reconstructions.

CHU Runs Analysis

Ranked Low Snowpack	Low Snowpack Period	Period Duration	Magnitude (running total/period)	Intensity (Magnitude/ Duration)	Ranked High Snowpack	High Snowpack Period	Period Duration	Magnitude (running total/period)	Intensity (Magnitude/ Duration)
1	1729-1742	15	-27.059	-1.804	1	1915-1932	18	155.276	8.626
2	1894-1914	22	-32.241	-1.465	2	1979-1988	11	107.893	9.808
3	1950-1977	28	-37.922	-1.354	3	1826-1840	15	133.521	8.901
4	1818-1829	12	-19.419	-1.618	4	1935-1945	11	105.977	9.634
5	1841-1848	8	-6.392	-0.799	5	1759-1774	16	134.387	8.399
6	1999-2006	8	-11.212	-1.402	6	1719-1727	9	82.486	9.165
7	1707-1713	7	-5.151	-0.736	7	1849-1859	11	93.878	8.534
8	1667-1676	10	-10.979	-1.098	8	1687-1695	9	78.017	8.669
9	1879-1887	9	-8.462	-0.940	9	1743-1751	9	80.453	8.939
10	1752-1758	7	-10.866	-1.552	10	1656-1666	11	88.921	8.084

WSR Runs Analysis

Ranked Low Snowpack	Low Snowpack Period	Period Duration	Magnitude (running total/period)	Intensity (Magnitude/ Duration)	Ranked High Snowpack	High Snowpack Period	Period Duration	Magnitude (running total/period)	Intensity (Magnitude/ Duration)
1	1818-1834	17	-59.921	-3.525	1	1718-1727	10	54.347	5.435
2	1728-1744	17	-59.908	-3.524	2	1978-1988	11	30.759	2.796
3	1893-1908	15	-58.176	-3.878	3	1791-1804	14	36.300	2.593
4	1950-1964	15	-40.028	-2.669	4	1783-1787	5	26.795	5.359
5	2000-2007	8	-22.915	-2.864	5	1907-1945	39	88.053	2.258
6	1751-1765	15	-31.283	-2.086	6	1866-1871	6	24.606	4.101
7	1773-1782	10	-26.083	-2.608	7	1762-1772	11	24.845	2.259
8	1860-1865	6	-17.364	-2.894	8	1883-1892	10	23.352	2.335
9	1841-1848	8	-18.939	-2.367	9	1833-1840	8	20.740	2.592
10	1967-1977	11	-21.652	-1.968	10	1743-1750	8	19.541	2.443

Figures

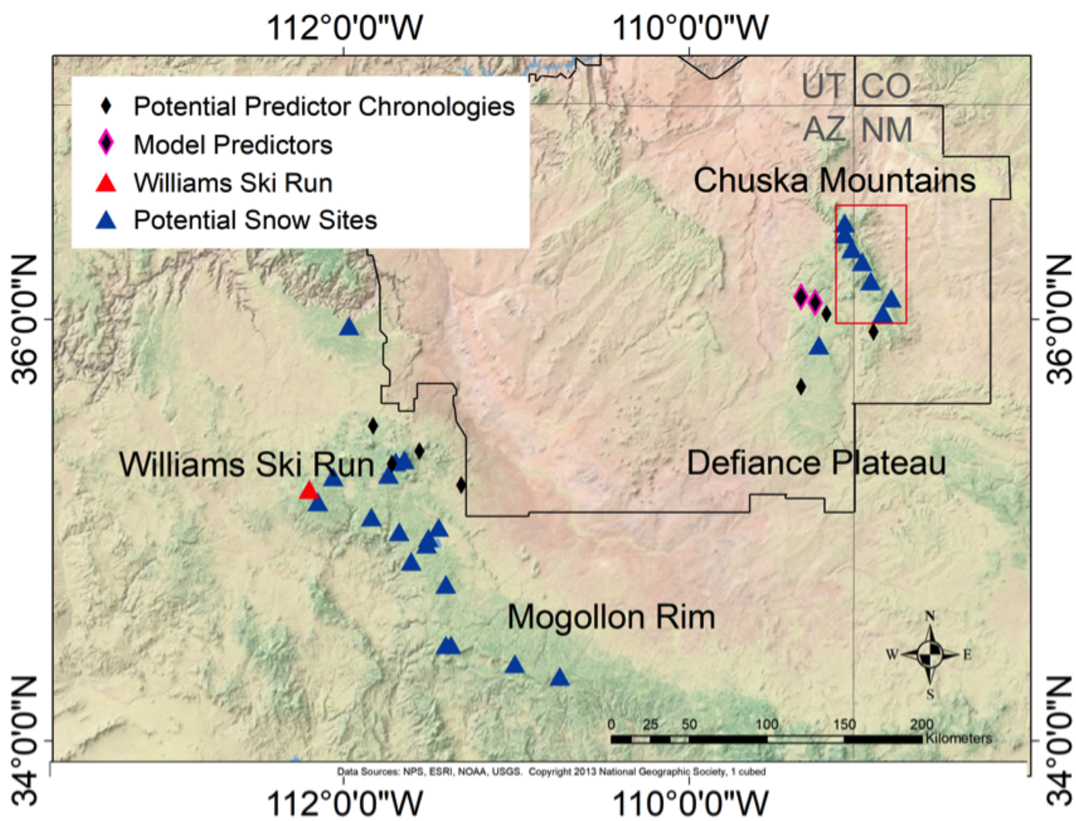

Figure 1 The Navajo region of northeastern Arizona and northwestern New Mexico. The Navajo Nation boundary is in black. The Four Corners (the intersection of the four states Colorado, Utah, Arizona and New Mexico) is in the upper right portion of the map. The location of snow sites and tree-ring sites used in this study are shown, blue triangles and black diamonds, respectively. (Spider Rock chronologies overlap at the map scale.) Model predictors are indicated as pink diamond outlines. The Williams Ski Run (WSR) model predictand is indicated with the red triangle. Snow sites used for the Chuska Mountains SWE predictand (CHU) are within the red box.

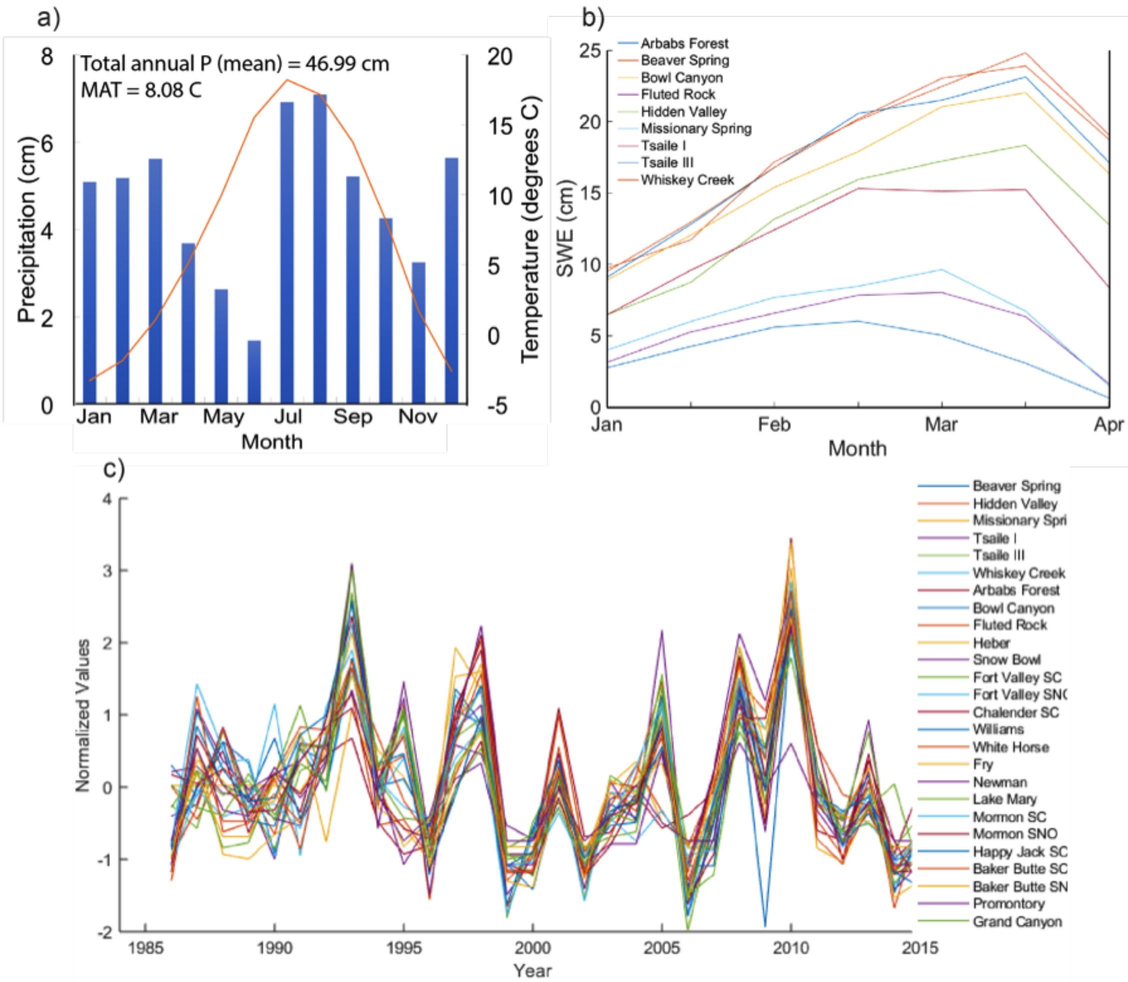

Figure 2 a) Climograph for the Navajo region using PRISM data (PRISM Climate Group, Oregon State University, http://prism.oregonstate.edu, created 20 May 2018). b) Chuska mountain snow sites average monthly SWE (1985-2016). c) March SWE from snow measurement stations in the Chuska Mountains, San Francisco Peaks, and Mogollon Rim, Arizona (common period 1986-2015).

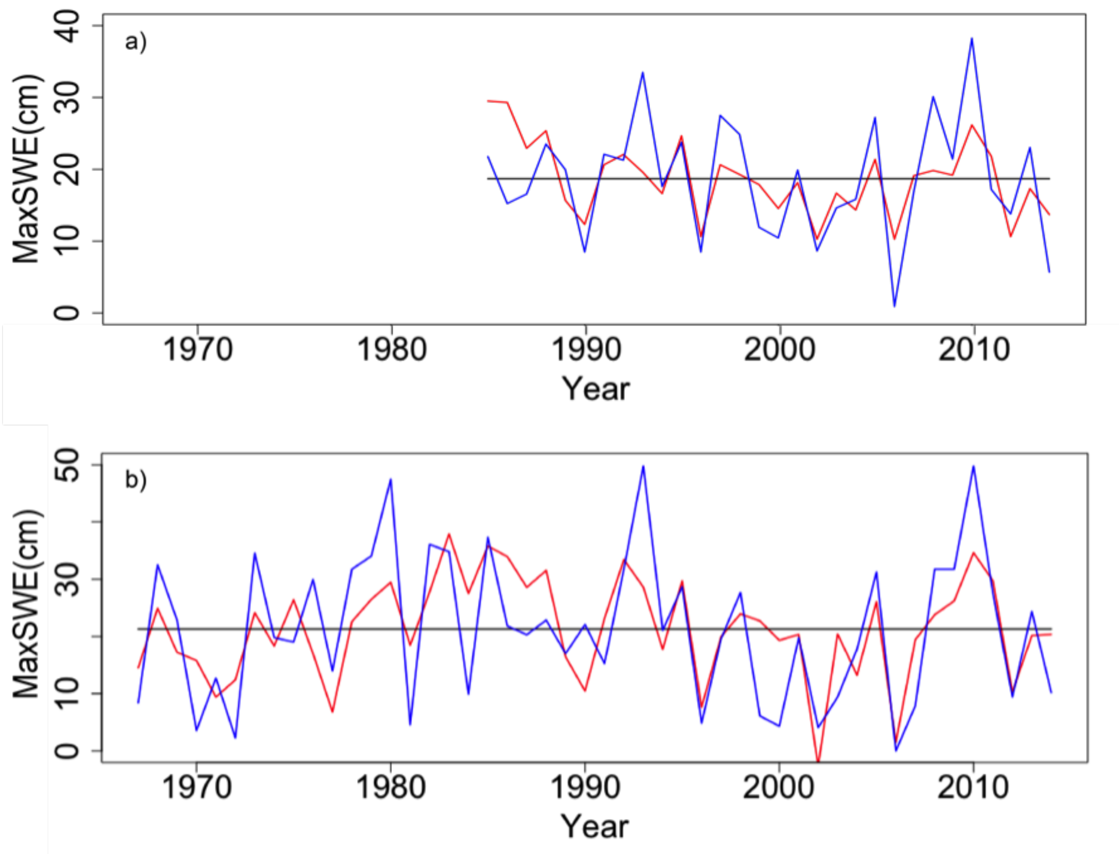

Figure 3 Observed (blue) and predicted (red) MaxSWE (Mar 1) at a) CHU Chuska mountain local snow for the years 1985-2014 and b) WSR Williams Ski Run for the years 1967-2014.

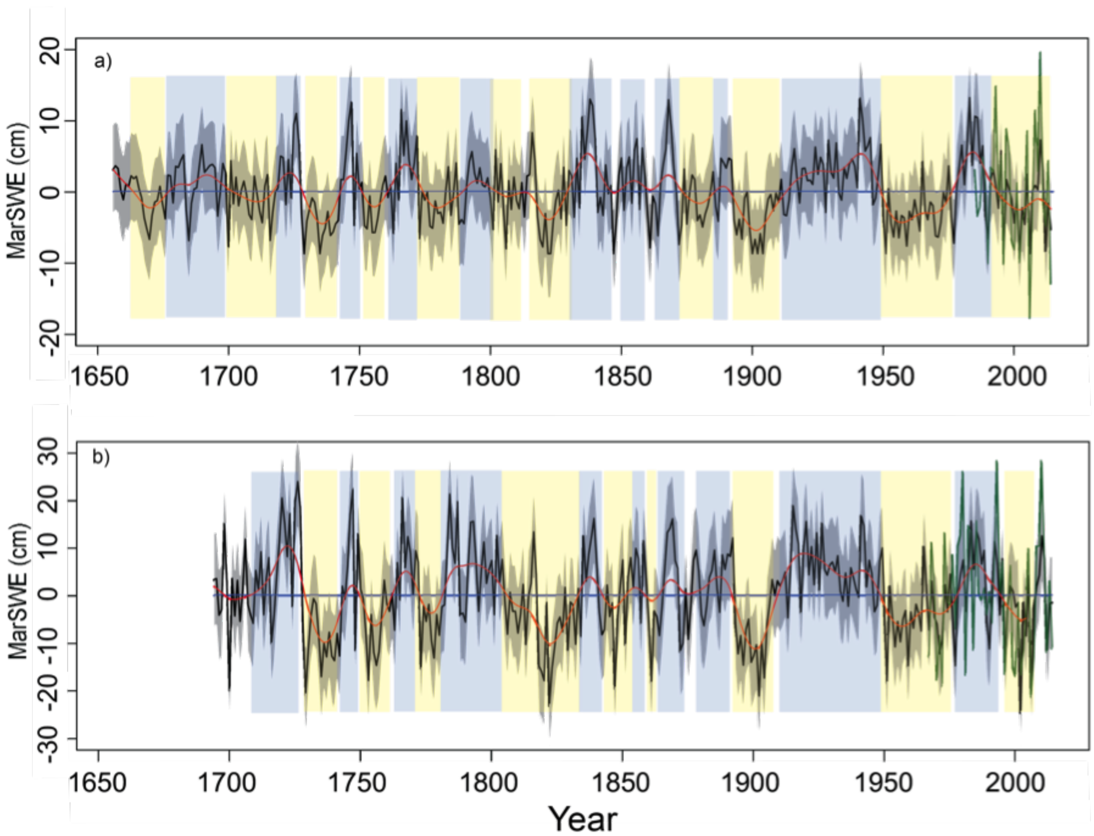

Figure 4 MaxSWE (March 1) reconstructions (black) with the 95% confidence interval in grey shading for a) CHU SWE (1656-2014 AD) and b) WSR SWE (1694-2014 AD). The zero mean is the blue line. The red line is the 20-year cubic smoothing spline. The green line is the calibration series. Multi-year periods of above and below average conditions are highlighted in blue and yellow, respectively.

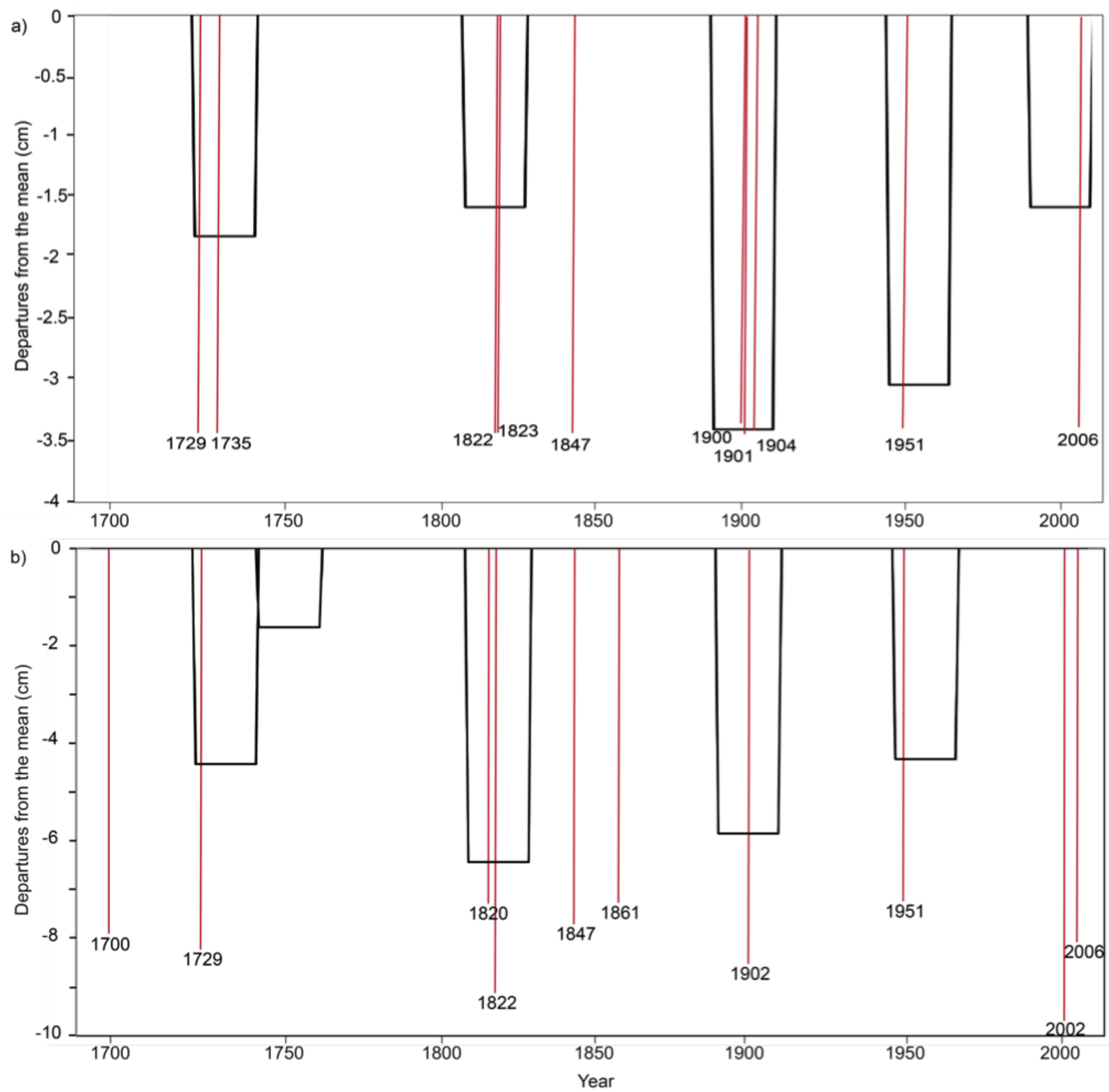

Figure 5 Snow droughts in the a) CHU Chuska mountain reconstruction and b) WSR Williams Ski Run reconstruction. Black boxes are the average deficit for the 5 driest 20-year periods. Red lines are the driest ten single years in each reconstruction.